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Summary

Mechanisms by which changes in the land surface may affect climate in lower latitudes are discussed and relevant general circulation model experiments are reviewed. Albedo and surface moisture availability appear to be the most important factors. Results are presented of experiments designed to assess the effects on rainfall of variations in model land surface properties in the Sahel such as could follow from a loss of vegetation in the region. With reduced maximum soil moisture, an extensive decrease of rainfall is obtained over the western Sahel but there are increases over the Sudan and Ethiopia. When, additionally, albedo is enhanced, the decreases extend to the Sudan and northern Ethiopia. In a series of experiments with changes of albedo and soil water capacity representing the deforestation of the tropics, major reductions of rainfall are obtained mainly due to increases in albedo. These reductions extend outside the deforested regions into the summer hemisphere though further experiments are needed to confirm that this is not a random difference due to model variability.

1. Introduction

There has been widespread concern about the impact on climate of changes in the African land surface, notably the losses of vegetative cover through the removal of forests in humid equatorial regions and of trees, shrubs, grass etc. by overgrazing in semi-arid regions. This concern has been heightened by the occurrence of droughts, notably that in the Sahelian region of Africa which has persisted since 1968 (e.g. Dennett et al, 1985).

In this paper general circulation model (GCM) experiments relevant to this problem are briefly reviewed (Section 2) and preliminary accounts are given of some recent experiments with the Meteorological Office model made to assess the effect on rainfall of removal of vegetation in the Sahel region (Section 3) and replacement of rain forests by grazing land (Section 4). Modelling studies show that the tropical rainfall distribution can also be strongly influenced by tropical ocean surface temperature (e.g. Rowntree 1972, Keshavamurty, 1982). The dependence of African rainfall on ocean surface temperatures is explored further by Folland, Parker and Palmer (1985).

2. Expected and modelled responses to variations in land surface properties

It will be useful, before reviewing results of experiments, to consider the likely effects on the energy balance of changes in the land surface in low latitudes. The discussion will be organised in terms of the processes and parameters by which the land surface is represented in models - albedo, longwave emissivity, soil moisture and surface roughness.

(a) Radiative

(i) Surface solar albedo

The albedo (fraction of incident radiation reflected) of a surface is of major importance in determining the net downward surface radiation. Large variations are possible due to the presence and variable characteristics of vegetative cover. Generally albedo decreases for a given vegetation type as the height of the vegetation increases, because of internal reflections, though for short sparse vegetation this may not be true if the albedo of the soil is low relative to that of the vegetation. Albedo also decreases generally for soil and vegetation as the surface wetness increases. The range of albedo for snowfree conditions is from about 0.1 for tropical forests to about 0.4 for some dry sandy surfaces. With incident mean daily solar fluxes typical of the tropics (300-400 Wm^{-2}

in cloudless conditions), a variation of about 100 Wm^{-2} is possible. However this would be an upper limit to spatial variations and is only conceivable at one place with an extreme climatic change or extensive human intervention – deforestation or irrigation on highly reflective soil. Actual changes on a large scale (1000 km upwards) seem unlikely to exceed about half this magnitude or 50 Wm^{-2}. These could be due to variations in surface wetness (Idso et al 1975, Norton et al 1979) or replacement of forests by grassland or dry soil.

Variations in surface albedo may be expected to affect local climate in two ways through their modification of the net radiation (R_N). Firstly, a reduction in R_N decreases the energy available for upward turbulent transfers of sensible and latent heat and for downward transfer into the soil (G). The partitioning of the decrease in (R_N - G) between sensible (H) and latent (LE) heat flux has been discussed by Rowntree (1984). For typical tropical conditions, the changes in LE will be a fraction $x = 3/(4 + r_s/r_{aE})$ of the change in (R_N-G) where r_s and r_{aE} are the surface and atmospheric resistances to evaporation. The value of x varies from about 0.75 for a moist surface to zero for a dry surface. Thus, as (R_N-G) decreases, the evaporation will decrease – for incoming solar radiation of 250 Wm^{-2}, a change in albedo of 0.1 will decrease (R_N-G) by 25 Wm^{-2} if G is unchanged, reducing evaporation by 0.65 mm/day if $r_s = 0$. Such a decrease is likely to lead to a reduction in rainfall as discussed later in this section.

The second mechanism by which surface radiative characteristics can modify tropical climate is that discussed by Charney (1975) whereby heating of a vertical column (surface and atmosphere) relative to adjacent regions can increase ascent and vice-versa. The geographical distribution of heat sources and sinks thus leads to vertical circulations which are mainly responsible for limiting the temperature gradients to those required in the tropics by the dynamical constraint of low rotation rates. Thus an increase in surface albedo, by weakening the heat source, tends to reduce the ascent and associated rainfall. This mechanism can be enhanced by drying of the atmosphere (which increases upward longwave radiation) and drying of the land surface which as discussed in section 2(b) can cause

surface heating and so also increase upward longwave radiation. These may be counterbalanced by a decrease in cloudiness reducing reflected solar more than it increases upward longwave radiation.

An impression of the possible impact of albedo changes is gained by noting that typically in the tropics the net radiative heating of the surface of ~ 150 Wm^{-2} is offset by an atmospheric radiative cooling of ~ 100 Wm^{-2}. With clear sky solar radiation of 400 Wm^{-2} a 0.1 increase in albedo will thus eliminate most of the net heating.

(ii) GCM experiments with surface albedo changes

For a more quantitative estimate of the effect of albedo changes we must consider the results of GCM experiments. However, in doing so it must be remembered that the complexity of the interactions between radiation, cloud and atmospheric and surface moisture content require a degree of realism in modelling surface and cloud processes which is probably not yet attainable. The GCM experiments made to date therefore cannot be expected to give more than an indication of the likely impacts.

Studies with GCMs on the effects of albedo variations, some of which include albedo changes in more than one area, are listed in Table 1. All are for northern summer. Except as indicated, the models included interactive surface hydrology and radiative transfer dependent on modelled water vapour and cloud.

Although the global-scale change in albedo studied by Carson and Sangster is not a realistic change, their results are instructive in showing clearly the major features of a model's response. In their first experiment (Fig. 1(a)) albedos of all snow-free land were increased from 0.1 to 0.3; the land was kept wet to avoid the confusion of soil moisture-albedo interactions. The major effects were:

(1) Higher surface pressures over land (by up to 12 mb in the inner northern continents)

(2) Decreased average evaporation over land (from 3.65 to 2.7 mm/day)

(3) Decreased average precipitation (and ascent) over land (from 4.6 to 3.4 mm/day) with compensating increases over oceans (from 2.9 to 3.25 mm/day).

Note that the decrease in precipitation over land exceeded that in evaporation, showing it was partly due to a decrease in moisture convergence. These results are consistent with the above discussion.

Similar results were obtained in a second experiment (Fig. 1(b)) in which soil moisture was interactive. In both experiments but especially the second, some land areas mostly on the western edge of continents (western North Africa, Europe) were wetter with the higher albedo probably because the higher pressures over the land mass to the east decreased the equatorward advection of dry air.

The other experiments in Table 1 all involved albedo increases over relatively small areas. The principal results (Table 2) are:

(1) Evaporation was decreased in all cases

(2) Precipitation was decreased in all experiments except two where changes were small; both these were near mountain ranges (Rockies, Himalayas) and poleward of latitude 20°.

(3) Moisture convergence was decreased in all but 3 cases, these 3 being poleward of latitude 20°.

Although, for Chervin's experiment, there is no direct evidence concerning (1) and (3), indirect evidence is provided by the maps showing decreased ascent and reduced soil moisture with increased albedo, especially over the Sahara.

For the areas equatorward of 20° latitude which are most relevant to this paper, the ratios of the fractional change in rainfall to the change

in albedo (last column of Table 2) average about -2 suggesting a 20% decrease in rainfall for a 0.1 increase in albedo. On average for the limited area anomalies, more than half the change is due to moisture convergence, though there is considerable variability in this, particularly between Charney et al's results and those of Sud and Fennessy.

An important aspect of the results is the effect of albedo on cloud amount. Here also Sud and Fennessy's and Charney et al's results differ, the latter finding decreases in cloudiness of 7 to 24% in 5 of their 6 cases, whereas Sud and Fennessy obtained decreases of at most 4%. In consequence radiative impacts were greater in Sud and Fennessy's experiment (30 Wm^{-2} in net surface radiation compared with 17), yet rainfall was less affected.

Carson and Sangster (as reported by Rowntree (1983)) made experiments with albedo dependent on soil moisture through a simple linear relation from 0.15 for a moist to 0.30 for a dry surface. This provides a positive feedback to any tendency for change in surface moisture and may be expected to enhance contrasts between wet and dry regions. This expectation was confirmed with sharpened contrasts between southern India and northwestern India/Pakistan and across the Sahel by Days 21-50; later however there was a northward expansion of rainfall over western North Africa, probably associated with the higher albedo over much of Africa and Asia for the reasons suggested above in connection with Carson and Sangster's albedo experiments. Laval (1983) has also recently run experiments with soil moisture dependent albedos in the LMD model, as well as with increased prescribed Sahel albedos; with the latter she noted a tendency for weakening of the upper tropospheric easterlies in the Sahel region.

iii. Surface longwave emissivity

The effect of changes in surface longwave emissivity for downward and upward radiation ($\delta\varepsilon_*$, $\delta\varepsilon_{*\uparrow}$) on the net surface radiation may be written:

$$\delta R_N = R_L \, \delta\varepsilon_* - \sigma_S T_*^4 \, \delta\varepsilon_{*\uparrow} \qquad (17)$$

where σ_S is Stefan's constant, R_L is downward longwave radiation and T_* is surface temperature. Typically in the tropics, $R_L \sim 0.85\ \sigma_S T_*^4$ (eg Viswanadham 1972) so that if

$\delta\varepsilon_* = \delta\varepsilon_{*\uparrow}$ and $\sigma_S T_o^4 \sim 450\ Wm^{-2}$ ($T_o \sim 27°C$),

$$\delta R_N \approx -67.5\ \delta\varepsilon_*\ (Wm^{-2})$$

A survey of data on surface emissivities was presented by Kondratyev et al (1982) who commented on the inadequate spectral detail of the available observations. Their limited data suggest that the emissivity integrated over the longwave spectrum is about 0.92 for soils and 0.94 for vegetation, while for the 9-12 μm 'water vapour window', the mean for dry soils is 0.96 and for vegetation 0.98. Whilst the relative uncertainties in these figures must be substantial, the suggested value of 0.02 for $\delta\varepsilon_*$ for both the integrated emissivity appropriate for upward radiation and the water vapour window which is more appropriate for the net longwave radiation is so small that one may have sufficient confidence in the consequent estimate of

$$\delta R_N \sim 1.35\ W\ m^{-2}$$

to assert that the effects of land surface changes on net radiation through emissivity changes can be neglected in comparison to the changes due to surface albedo. The only relevant GCM experiment known to the writer is that reported briefly by Hansen et al (1983), who found "little effect on the general circulation or atmospheric temperatures" from the introduction of spectrally dependent emissivities for deserts, snow and ice.

(b) Response to surface moisture availability

(i) General considerations

The ground hydrological processes determine, in conjunction with the characteristics of the soil and vegetation, the surface moisture availability which controls the partitioning between sensible and latent heat fluxes. In moist conditions, of order 80% of this quantity is

transferred in the latent form so that the change from moist to arid conditions can enhance the direct heating of the atmosphere by about 100 Wm^{-2}. This gain may of course be compensated by a loss of latent heating if precipitation is affected by the lack of evaporation. There is ample evidence that this occurs as discussed later.

A calculation of the effects on relative humidity of such repartitioning is instructive. If an evaporation rate of 3mm/day is reduced to 1mm/day by limited surface moisture availability the effect on the tropospheric mean relative humidity of the repartitioning (including 58 Wm^{-2} more sensible heat flux) is about 12%/day. Clearly such a large difference could affect the preciptiation from an initially moist air mass within a day or less. It is interesting that the increase in saturation vapour pressure due to the warming contributes nearly half the decrease in relative humidity. Mintz (1984) has discussed the summer moisture budget over the eastern United States to demonstrate the important role of local evaporation. He suggests that because surface evaporation moistens the boundary layer it can stimulate convective instability which leads to the condensation of moisture advected from the ocean as well as that evaporated from the land.

(ii) GCM experiments

A review of GCM experiments with global changes in soil moisture and evaporation over land has been prepared by Mintz (1984) and discussed further by Rowntree (1984). Generally there is a clear decrease in precipitation when evaporation over land is reduced. The effect on African rainfall at 15°N in these experiments is summarized in Table 3; all the experiments show a decrease in rainfall with reduced evaporation.

Of more relevance here are experiments with limited area soil moisture anomalies. There have been two model experiments with soil moisture anomalies over a region similar to the Sahara in summer, one with a cyclic channel model (Walker and Rowntree 1977), one with the Meteorological Office 11-layer global model (Cunnington and Rowntree 1985). The effects of dry anomalies may be summarized as follows:

(1) Evaporation is decreased (from about 3 to 0 mm/day).

(2) The surface warms (by ~ 20K) and sensible heat flux increases (from about 30 to 120 Wm^{-2}).

(3) The lower layers (up to about 600mb) warm, with cooling (~ 2K) at higher levels.

(4) Surface pressure decreases while at 700 mb and above, geopotentials are higher with a more anticyclonic circulation.

(5) Precipitation decreases (from ~5 to <1 mm/day) so that the initial anomaly in soil moisture persists.

(6) Ascent over the anomaly becomes limited to a shallow layer so that it is insufficient with dry air to give significant rainfall.

It should be noted that the radiative schemes in these experiments were relatively simple. Walker and Rowntree prescribed constant surface heating and atmospheric cooling rates so that none of the feedbacks discussed in section 3(a) could operate. Cunnington and Rowntree used a more complete radiative calculation but still prescribed clouds. The impact of moistening of the soil on the surface radiation budget was considerable. Averaged for days 7 to 16 the net radiation was 110 Wm^{-2} with initially dry soil and 164 Wm^{-2} with initially wet soil. With model-dependent cloud, this contrast would probably have been reduced and the persistence of the anomaly could have been less marked.

Sud and Fennessy (1984) studied the effects of eliminating evaporation in the Sahel region (12° – 20°N). Averaged for the last 30 days (July) of a 47 day integration, they obtained a small (0.4 mm/d) increase of precipitation in association with increased moisture convergence into a region of lower surface pressure over the evaporation anomaly. The control experiment's rainfall is excessive both over the anomaly (5.6 mm/d compared

with 3.5 mm/d observed) and further north with more than 1 mm/d over almost the whole Sahara; there is associated excessive evaporation over the Sahara (Shukla et al, 1981). The presence of a spurious moisture source to the north may have contributed to the maintenance of rainfall over the anomaly.

Another relevant experiment is that of Yeh et al (1984) using the version of the GFDL model with idealised geography. They enhanced the initial soil moisture over 30° wide latitude bands and studied the model's response over the next five months. The precipitation over land was significantly increased not only over the initial anomaly but, in some cases, in adjacent regions - e.g. near 15°N for an anomaly in 30°-60°N in northern summer and near 10°S for an anomaly at 0°-30°N in southern summer. Similar effects were found with a European soil moisture anomaly by Rowntree and Bolton (1983).

(c) Response to variations in surface roughness

The effects of variations in surface roughness have been studied little with GCMs. For given atmospheric conditions (wind, humidity, etc) it is possible to show (Rowntree (1984), see also Dickinson (1980)) that the effect of an increase in roughness on partitioning between the sensible and latent heat fluxes depends on the available energy ($R_N - G$), the humidity and the surface and atmospheric resistances to evaporation. Increasing the surface roughness (decreasing the atmospheric resistance) increases evaporation for low surface resistance (e.g. a wet surface) and decreases it for high surface resistance. For typical daytime conditions in the tropics, the critical value of surface resistance at which roughness changes have little effect on the partitioning is near 50 sm^{-1}, which is below that typical of much vegetation which has no surface water and has transpiration limited by soil moisture availability. Thus during dry spells forests may give less evaporation than grassland if the stomatal resistances are similar and larger than the critical value, the converse of the difference in wet conditions.

3. Soil Moisture storage and drought in the Sahel

(a) Introduction

The Sahel region of Africa has been subject throughout the twentieth century to increasing pressures, of population, both human and animal (USA Nat Acad Sci, 1983), and more recently of drought. After a period of relatively ample rainfall during the 1950s, deficient precipitation set in during the 1960s and since 1968 every year has had below normal rainfall, as measured by regional means for the western and central Sahel (about 11° - 18°N, 20°W - 20°E) (Lamb 1983, Dennett et al 1985, Gregory 1982). Generally dry conditions affected areas both north and south of the Sahara, extending to beyond the equator (Nicholson 1983a).

The possible role of positive feedbacks in the persistence of the Sahel drought has been widely discussed (USA Nat Acad Sci 1983, Hare 1983). One proposal (Charney 1975) is that increases in albedo could provide such a feedback; these occur as vegetation dries or decays, revealing increasing areas of generally highly reflective dry soil or sand. Another (Walker and Rowntree 1977) is that decreases in evaporation could lead to a drier atmosphere and reduced rainfall. Other proposed mechanisms, requiring further investigation, involve ocean surface temperatures (e.g. Lamb 1978) and dust (e.g. Rasool 1984).

Modelling studies have given consistent support for the albedo mechanism (Section 2(a)) though evidence (Norton et al 1979, Courel et al 1984) for actual albedo change is both meagre and inconclusive. A dependence of rainfall on evaporation has been found in several experiments (Section 2(b)), though often the anomaly in evaporation has been applied on a larger scale than that of the Sahel. For anomalies of smaller scale, it appears possible for the increase in direct heating of the atmosphere associated with decreased evaporation to lead to increased rainfall (Sud and Fennessy 1984). Though relevant observational data on evaporation are lacking, it is obvious that for arid and semi-arid regions where evaporation is limited by rainfall, the drought must have been accompanied

by a decrease in evaporation. Also, as discussed by Nicholson (1983b), though little direct observational data exist on the occurrence of these feedbacks, the very persistence of the drought provides indirect evidence.

To clarify the issue of the effect of reduced evaporation, we have run two experiments with decreased evaporation over the Sahel. Rather than eliminate evaporation entirely, we have attempted to come a little closer to reality by reducing the capacity of the soil to hold water accessible to the atmosphere for evaporation from a value of 15 cm, appropriate for deep rooting vegetation such as trees and permanent grasses, to 1 cm, more appropriate for bare soil. Though lower than the 2 to 3 cm often assumed, for the Sahel the estimate for bare soil is supported by observations in Niger by the Reading University TAMSAT project showing that after irrigation evaporation was near the potential rate for only about one day and after 5 days with a cumulative evaporation of about 1 cm, evaporation rates had become small (< 1 mm/day).

The experiments described here may be regarded as a sequel to those of Cunnington and Rowntree (1985) showing marked sensitivity of the modelled Saharan July circulation and rainfall to changes in atmospheric and surface moisture content. These demonstrated that a global model could, with an initially wet (dry) Sahara generate a circulation which maintained this wet (dry) condition for at least 50 days. It was found that rainfall may be inhibited by absence of soil water storage. This is a similar result to those of Shukla and Mintz (1982) and Sud and Arakawa (reported by Mintz (1984)) though in those experiments the suppression of evaporation was on a global scale. Although total inhibition of evaporation is not possible, the condition may be approached with bare soils such as may be produced through overgrazing in the Sahel region (Mensching 1980). The lack of vegetation prevents interception (and subsequent evaporation) of the rainfall, so reducing the rapid short-term feedback of moisture to the atmosphere which is important in the moist tropics (Shuttleworth et al, 1984). With no roots, much of the water that does infiltrate into the soil can pass too deep for subsequent diffusion to the surface, either raising the water table or leaving a deep moist layer of soil or sand which can persist for years (Cloudsley-Thompson, 1977, p.12). Where crusting of the

surface occurs due to drying out after rain, a larger part of the next rainfall runs off (Mensching 1980). Except when dormant vegetation can take advantage of the occasional heavy rainfall (e.g. through seed germination), moisture will only be readily available for evaporation for the day or two after rain before the surface dries. In the Sahel, rainfall is modulated by easterly waves which have a period of about four days (Burpee 1974, Reed 1978). This period is long enough for the surface soil layer to dry out, so that, in the absence of vegetation to tap the deeper soil, surface evaporation may provide little of the moisture needed for rain (see also Mensching 1980, Perrier 1982, p. 402, Warren and Maizels 1977).

It may be argued that after heavy rainfall, annual seeds will germinate and, in subsequent growth, tap moisture at greater depths. However, in parts of the Sahel, it appears that degradation of vegetation and soil may be so severe that the nutrients needed for vigorous growth may be restricted by soil erosion and by leaching in occasional heavy storms (USA Nat. Acad. Sci. 1983, Warren and Maizels 1977); if so, associated root growth may be too shallow for such plants to exploit significantly more moisture than would be evaporated from bare soil. Even where deeper-rooting vegetation does exist or develop, it is liable to be patchy, covering only a small fraction of the surface, because of the uneven character both of the infiltration and, in part as a result, also of the root systems. Unfortunately, there appears to be a dearth of relevant published data on these matters so that it is very difficult to assess how much the combination of drought and human and animal pressures on the land have actually affected moisture availability. It should thus be useful to take the inverse approach to the problem: ie include such land surface anomalies in a model and assess how the model's response to these compares with the observed atmospheric anomalies.

(b) Model and experiments

The model used is a development of that used by Cunnington and Rowntree (1985) and Slingo (1985). The horizontal grid mesh has been changed from a quasi regular 220 km mesh to a $2^1/_2$ by $3^3/_4$ latitude-

longitude grid. The long wave radiation scheme has been improved (Slingo and Wilderspin 1985) and a geographical distribution of surface albedo included (Wilson 1984). The model uses a year of 360 days, consisting of twelve 30-day months (Slingo, 1982). The formulation of evaporation (E) is now

$$E = \beta E_p$$

where the potential evaporation E_p allows for a typical surface resistance of 60 s m^{-1}. Dependence on soil moisture (m) is introduced through

$$\beta = m/m_c, \text{ with } \beta = 1 \text{ if } m > m_c$$

where m_c is a critical value of m, taken as 5 cm in this model. Runoff to rivers or ground water occurs only if m exceeds the maximum accessible water storage (m_{max}) which, as discussed earlier, is taken as 15 cm in the control experiment and is reduced to 1 cm in each of two anomaly experiments over 10°-32$^1/_2$ N, 20°W-60°E - ie Africa and Arabia north of 10°N and semi-arid regions of southwest Asia. Over most of this area, the change is of little consequence because soil moisture seldom exceeds 1 cm at 20°-32$^1/_2$°N in the relevant periods of the control integration. The effect of the change on maximum evaporation rate is less than might be expected because as m decreases below m_c, E_p increases. Thus E is not proportional to β but, as shown previously (Rowntree and Bolton 1983) approximately to $\beta^{1/2}$. Thus, by limiting m to 1 cm, we are restricting E to $(0.2)^{1/2}$ or about 45% of its value with m > 5 cm. This may be regarded as assuming rainfall to be so distributed that not more than 45% of a grid box has evaporation at the potential rate at any one time, with little evaporation elsewhere; this is a reasonable assumption in a region of mostly convective rainfall.

Two experiments with m_{max} of 1 cm were run, starting from March 1 of Years 2 and 3 respectively of a 4$^1/_2$ year control experiment which started at July 24 of Year 0. Both were run for 270 days, terminating at the end of November and results were compared with the means for Years 1 to 4 of the control run. Assessments of statistical significance were made using

Student's t-test with the null hypothesis that the 4 control and 2 anomaly years were from the same population. The boreal summer (June - August) and autumn (September -November) rainfall distributions for the control run (Fig. 2) are in good agreement with observations over Africa apart from excessive rainfall over Ethiopia and Somalia in autumn.

(c) Results

Figure 3 shows the effects of reducing m_{max} in June-August; both experiments show a reduction of over 1 mm/day extending from near 10°N 25°W across Senegal and southern Mali to Niger. Most of this area is significant at the 10% level on Student's t-test (Fig 3(b)). Detailed maps (not shown) indicate significance levels of below 5% over much of the area and 1% at a few points. Parts of Chad and western Sudan at 12°-17°N are also affected though less homogeneously than further west. The increases near 15°N 15°E appear to be a chance phenomenon associated with a minor minimum in the control run not present in the anomaly experiments. However the increases further east are systematic, appearing in most monthly means of both experiments. These increases are associated with increased poleward flow of moist boundary layer air on the eastern side of a fall of pressure (Figure 4) caused by the warming of lower tropospheric air (Figure 5) over the area with reduced soil moisture. This warming is familiar from earlier experiments (e.g. Walker and Rowntree 1977); with less moisture available, the heat received at the surface through radiative processes is transferred to the atmosphere increasingly in the form of sensible heat rather than latent heat, so warming (and drying) the lower troposphere. The changes south of 10°N are discussed later.

The monthly differences show that in May and June there is no consistent reduction in rainfall over the anomaly area (a result similar to that of Sud and Fennessy (1984)) whereas August and September both have well-defined decreases.

In September-November (Fig 6) there is very good agreement between the two experiments over most of the area shown. The decreases over the area with decreased m_{max} are largest at 0°-15°W; the decreases extend south of

the equator at 10°-25°E. Again increases are dominant east of 30°E especially in the first experiment. There are well-defined changes over the Atlantic as the inter-tropical convergence zone shifts southward. Apart from this feature, which appears in August, September and October for each experiment, the seasonal difference maps exhibit little consistency outside equatorial and northern Africa.

(d) Discussion

In assessing the relevance of these results to the Sahelian drought, we should consider the realism of the perturbation applied to the model, and how the timing, extent and magnitude of the impact on model precipitation compare with the observed rainfall anomalies. We have already discussed the first question and concluded that at present lack of data precludes a proper assessment. A quantitative study is needed, for a representative sample of locations, of the past and present vegetation with some method of interpreting the data in terms of moisture availability to the atmosphere. Until this is possible it is necessary to resort to verification of the impacts on measurable quantities such as rainfall. This is not really satisfactory both because other changes (eg albedo) may cause similar impacts and because, on the other hand, deficiencies in the model may prevent accurate simulation of the impacts, so falsely giving the impression that the tested mechanism is not relevant.

Concerning timing, the downward trend of Sahel rainfall has been most marked in August and September with no significant trend in June (Dennett et al 1985). Compared with 1931-60, their data show 1968-83 mean anomalies of -8, -13, -27 and -23% for June, July, August and September respectively. A similar change in character between early and late summer is shown by Gregory's (1983) geographical analysis of changes from 1931-60 to 1961-80 and is also evident in our results.

The magnitudes of the observed and modelled differences also appear similar. In the driest years (1972, 73, 82, 83) Dennett et al (1985) found a June–September deficit of 33–38%, which near 14°N is equivalent to about 1.5 mm/day, similar to the June–August impacts at 15°W–10°E in our experiments.

The spatial extent of the annual anomalies has been studied by Nicholson (1981, 1983a). For 1971–73, regional mean anomalies were negative over all of Africa from 10°S to 30°N; normalised by the local standard deviations, they were largest at about 12°–18°N west of about 27°E. There is no evidence in Nicholson's analyses for anomalies of opposite sign east of about 30°E such as our experiments suggest. The changes south of 10°N in the model are of particular interest since these are outside the region of reduced m_{max} suggesting that equatorial rainfall at this season depends in part on moisture evaporated north of 10°N. A similar latitudinal expansion was obtained by Yeh et al (1984) as discussed in Section 2(b). West of 15°E the modelled rainfall is increased in June–August and decreased in September–October south of 10°N. Nicholson's (1981) cross sections based on monthly anomalies for 1971–73 at 0° longitude suggest similar features. At 15°–30°E, rainfall south of 10°N is decreased in both seasons in the model. Nicholson's (1983a) analysis of annual rainfall for 1968–73 shows that the negative anomalies extended to the equator and her cross sections for 26°E (Nicholson 1981) at 5°–10°N for 1971–73 (no data south of 5°N) suggest the decrease occurred in most months.

Summarising, the response of the model's rainfall to a reduction north of 10°N over Africa in available moisture storage appears similar to the observed anomalies of recent decades over Africa north of the equator west of about 25°E, but not to the east of that longitude. This result could imply that there is another mechanism involved; this may be the albedo mechanism proposed by Charney (1975), which we have deliberately omitted here. An increase in albedo would be expected to weaken the heat low induced in the surface pressure pattern by the soil moisture changes and so reduce the east–west contrast in rainfall changes, as well as contributing further to a general decrease in rainfall.

To investigate this possibility, an experiment is being run identical to the second experiment discussed in this section except that semi-desert albedos are used in 10°-17$^1/_2$°N over Africa. Average albedo increases for the region affected (15°W - 48$^3/_4$°E) are .018, .071 and .022 at 11$^1/_4$, 13$^3/_4$ and 16$^1/_4$°N respectively. Figure 7 shows the June-August precipitation changes for this experiment.

4. Modelled impact of tropical deforestation

Land clearance for farming in areas of existing rain forest appears likely to affect the climate at least locally. The energy (R_N-G) available for turbulent fluxes will be reduced by an increase in albedo from rain forest values of about 0.12 to the 0.15-0.25 typical of pasture and partly cultivated land (Kondratyev et al, 1982). Partitioning of (R_N-G) between sensible and latent heat fluxes will be affected by changes in surface hydrology; this topic is discussed at length by Dickinson (1980). Latent heat flux tends to be greater from forests because (a) part of the rainfall is intercepted and subsequently evaporated at the potential rate (Shuttleworth et al, 1984), (b) maximum infiltration rates are generally higher, particularly in comparison with cropland and bare soil (Dickinson 1980, Sloan and Moore, 1984) and (c) the greater root depths, at least relative to bare soil and many crops, provide access to a greater depth of water so that evaporation will more readily continue through dry periods. There are also factors tending to reduce evaporation such as increased stomatal resistance in certain conditions (Perrier, 1982) and the larger roughness length in the conditions discussed in Section 2(c). Observational evidence of changes in run off associated with afforestation and deforestation (Dickinson, 1980, Oyabande, 1984) suggests that the factors acting to increase evaporation are generally dominant.

Because of the relatively simple formulation of surface hydrology in the Meteorological Office GCM (and most others), only two of these effects of forests, the lower albedo and greater available water capacity, were included in the deforestation experiments by Wilson (1984). Figure 8 shows the increases in albedo applied for the experiments M8 and M9 representing

the deforested (tropical grazing) state (M8 and M9 are called DEFOR and DEFALB by Wilson) relative to experiments L6 (ANNALB) and L7 (AFFHYD). They range from 0.04 to 0.10 depending on soil colour and whether the original vegetation type was woodland or forest. Figure 9 shows the reductions in available water capacity (m_{max}) applied in experiment M8 relative to L7. For free-draining soils, these are 12 cm (reduction from 18 to 6), 20 cm (32 to 12) or 32 cm (48 to 16) depending on soil texture. At points with impeded drainage the lower values were used in both experiments. For points which are not rain forest in the vegetation datasets, and for all points in L6 and M9, m_{max} was set to 15 cm.

Each of these experiments was run for 9 months from December 1 of Year 0 of the second annual cycle experiment with the Cyber version of the 11-layer model; this experiment was started from real data at July 24 of Year 0. Most of the results presented in this brief survey are for (M9-L6), representing the effects of the albedo change only. The signs and magnitudes of the differences in precipitation over the anomalies were generally similar in (M8-L7). Variations between (M9-L6) and (M8-L7) appear to be mainly due to chance except for the Amazon forest region later in the experiment when the greater water storage in L7 enhanced evaporation and rainfall. Some results for South America as well as Africa will be discussed here since they may be expected to be similar in character and the sample size is small. Interaction between the two regions is believed to be slight relative to the local impacts.

The differences in rainfall for (M9-L6) (increased albedo), averaged for the first 3 months (Fig 10) show general decreases exceeding 1 mm/d over both the Amazon and Congo forest regions. An interesting feature is the tendency for the rain belts which extend southeastward from the major rain areas over S America, Africa and Indonesia, the three regions where albedos are increased, to shift eastwards, giving belts of decreases over eastern South America, southeast Africa and eastern Australia with increases immediately to the east. These changes are especially clear in January and February (not shown) but not thereafter. This may be partly due to the seasonal cycle, but a higher noise level associated with the

development of differences in the land surface fluxes outside the anomaly regions (eg due to the changes in rainfall shown in Fig 10) may have contributed.

To allow a more quantitative assessment of the local changes due to deforestation, averages were calculated for each month for the Amazon basin region of South America and Congo basin region of Africa, each containing at least 60% of the points in their respective continents subject to deforestation and only a few other points (Table 4). Table 5 shows mean rainfall for these regions for experiment L6 and the changes (M9-L6) in rainfall and evaporation due to increasing albedo. Over the first two months the rainfall decrease considerably exceeds the evaporation decrease indicating that the Charney mechanism is leading to reduced water vapour flux convergence as in most albedo experiments (Table 2). Later, evaporation is constrained by rainfall so that changes in the two become similar.

The 9-month mean rainfall decrease of 0.9 mm/d over the Amazon region is somewhat larger than the annual mean decrease of 0.6 mm/d obtained by Henderson-Sellers and Gornitz (1984). This may be due to their inclusion of modelled cloud which appeared to act as a negative feedback though they also included changes in hydrology which may have amplified the response.

Table 5 also shows that the fractional rainfall decrease is greater over the Congo region, despite the larger albedo change over the Amazon region (Table 4). Two possible reasons for this can be suggested. During northern winter the decreases over Africa are compensated by large increases over the western Indian Ocean (Fig 10). It is possible that with the cooler tropical oceans around South America - both the east Pacific and Atlantic are generally cooler than the Indian and West Pacific Oceans - a larger reduction in land heating is needed to allow a major transfer of rainfall from land to sea, such as was a feature of Carson and Sangster's experiments (Fig 1). It is also possible that, because the African deforestation region occupies a smaller fraction of its continent than does

that in South America, rainfall is more readily diverted to other parts of the continent; this appears to have happened during northern spring (Fig 11(b)).

In the context of Sahel drought, the decreases over Africa near 10 N in July and August (Fig 11(b)) are to be noted. Fig 12(a) shows that this feature spans almost the full breadth of Africa in the June to August mean. This extension of anomalies into the summer hemisphere is reminiscent of the responses over both S America and Africa in southern summer (Fig 10). However it is a much less important feature in the experiment with anomalies in both albedo and maximum soil moisture (Fig 12(b)), and further experiments would be needed to confirm if it is a consistent effect of the land surface anomalies rather than model natural variability. The same remark applies to several other apparently rather random features of Fig 11(b).

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TABLE 1: LIST OF EXPERIMENTS ON EFFECT OF SURFACE ALBEDO CHANGES

Reference	Area	Latitude	Averaging period	Albedo Control	Albedo Modified	Comments
Carson and Sangster (1981)	Global	-	90 days	0.1	0.3	1,2,4
				0.2	0.3	1,2
Charney et al (1977)	Sahel	12-16N	31 days	0.14	0.35	4
	NW India	24-32N				
	Great Plains	32-48N				
	Central Africa	8-12N				
	Bangladesh	20-28N				
	Mississippi	32-48N				
Sud and Fennessy (1982)	Sahel	12-20N	31 days	0.183	0.3	
	NW India	24-32N		0.15	0.3	
	NE Brazil	4-24S		0.091	0.3	
	Great Plains	32-48N		0.129	0.3	
Chervin (1979)	Sahara	$7^1/_2$-$37^1/_2$N	60 days	.08-.35	.45	
	W USA	$27^1/_2$-$52^1/_2$N		.07-.17	.45	
Henderson-Sellers and Gornitz (1984)	Brazil	15.6S-7.8N	5 years	.11	.17	3,5

Comments

1. No moisture feedback in radiation.
2. No cloud feedback in radiation.
3. Roughness length and ground water capacity also changed.
4. Evaporation set equal to potential evaporation.
5. Ocean surface temperatures and sea ice not prescribed.

TABLE 2: EFFECTS OF ALBEDO CHANGES

Experiment	Change in Albedo ($\delta A*$)	Evap (δE) (mm/day)	Rainfall (δR) (mm/day)	$\delta R/R^{\neq}$	$\frac{\delta R/R}{\delta A*}$
Charney et al					
Sahel	.21	+0.9	+3.4	+.46	+2.2
NW India	.21	+0.5	+2.6	+.53	+2.5
Great Plains	.21	+1.0	+1.5	+.41	+2.0
Central Africa	.21	+0.7	+3.1	+.62	+3.0
Bangladesh	.21	+0.2	0	0	0
Mississippi	.21	+1.6	+1.1	+.25	+1.2
Sud and Fennessy					
Sahel	.12	+0.7	+1.5	+.26	+2.1
NW India	.15	+0.2	+0.5	+.13	+0.9
NE Brazil	.21	+0.3	+0.5	+.24	+1.1
Great Plains	.17	+0.2	0	.01	0.1
Chervin					
Sahara	.22		+2.5	+.4	+1.8
W USA	.33		+1	+.15	+0.5
H-S and Gornitz					
Brazil	.06	+0.45	+0.6	+.12	+2.0
Carson and Sangster					
0.1 0.3	.2	+0.95	+1.22	+.27	+1.3
0.2 0.3	.1		+0.4	+.14	+1.4

$^{\neq}$ Note: R is rainfall in the control experiment.

TABLE 3: CHANGES IN AFRICAN RAINFALL AT 15 N IN EXPERIMENTS WITH LARGE-SCALE REDUCTIONS OF EVAPORATION

Reference	Area	Anomaly Change in β^+ (wet dry)		15 N African rainfall (mm/d) Wet	Dry	Change
Manabe (1975)	Global	1	*	5	1	-4
Kurbatkin et al (1979)	Global	*	0	3	2	-1
Shukla and Mintz (1982)	Global	1	0	2-8	1-5	-2
Suarez and Arakawa (see Mintz 1984)	Global	1	0	7	1-2	-5
Carson and Sangster (1981)	Global	1	0≠	1-10	1-5	-3
Cunnington and Rowntree (1985)	Africa north of 10 N	1	0≠	2-7	0-2	-3

Notes: + $\beta = E/E_p$

* β is time-dependent function of soil moisture

≠ β is time-dependent function of soil moisture, the values given being the initial values

NB. All the rainfall data given are estimates from maps.

TABLE 4: AREAS USED FOR ANALYSIS OF DEFORESTATION EXPERIMENTS

Name	Latitude	Longitude	(a)	(b)	(c)	Albedo change
Amazon	15 S–5 N	75 –52$^1/_2$ W	48	42	21	.068
Congo	5 S–5 N	11$^1/_4$ –30 E	20	18	12	.051

Notes: (a) is number of model grid points in area
(b) is number of 'forest' points in area
(c) is number of 'forest' points for the same continent outside area

TABLE 5: MONTHLY MEAN RAINFALL (R) FOR EXPERIMENT L6 AND CHANGES IN RAINFALL (δR) AND EVAPORATION (δE) WITH INCREASED ALBEDO (M9-L6) FOR AREAS DEFINED IN TABLE 4

(Unit mm/d)

	Dec	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Ave
Amazon										
R	6.4	7.3	6.7	8.2	5.9	4.1	2.5	1.7	1.7	5.0
δR	–1.4	–1.3	–0.7	–2.4	–0.7	–1.5	–0.4	+0.2	–0.1	–0.93
δE	–0.6	–0.7	–0.7	–0.8	–0.7	–0.4	–0.4	+0.0	–0.0	–0.49
Congo										
R	4.3	4.9	3.4	3.8	7.5	3.8	1.4	.7	.7	3.4
δR	–0.5	–2.2	–2.3	+0.8	–1.2	–0.3	–0.8	–0.4	–0.3	–0.79
δE	–0.3	–1.1	–2.1	+0.3	–0.6	–0.4	–0.4	–0.7	–0.4	–0.72

Figure Captions

Figure 1 (a) Excess of precipitation with albedo of 0.1 relative to that with albedo of 0.3 at Days 21-110 (soil moisture fixed at 15 cm) in experiments with the Meteorological Office low resolution 5-layer model. Areas with decreases shaded.
(b) As (a) but for precipitation with albedo of 0.2 relative to that with albedo of 0.3 with interactive initially zero soil moisture.

Figure 2 Precipitation (mm/d) averaged for (a) June to August, (b) September to November of Years 1 to 4 of the control experiment which started on July 24 of Year 0. Contours at 1,2,5, 10 mm/d. Areas with over 5 mm/d shaded. The area with reduced moisture capacity is indicated in Fig 2(a) (dashed).

Figure 3 Differences in rainfall (mm/d), relative to the control average (Fig 2) for June to August of experiments with reduced soil water capacity starting from March 1 of (a) Year 2 and (b) Year 3 of the control experiment. Contours at 0, ±1,2,5 mm/d with increases (decreases) of over 1 mm/d hatched (stippled). Dashed lines in (b) outline the area significant at the 10% level with a 2-tailed Student's t-test.

Figure 4 As Figure 3(b) but for sea level pressure (mbar).

Figure 5 As Figure 3(b) but for 850 mbar temperature (K).

Figure 6 As Figure 3 but for September to November.

Figure 7 As Figure 3(b) for experiment with reduced soil moisture capacity and increased albedos.

Figure 8 Changes (M8-L7, M9-L6) in albedos imposed to represent deforestation of tropics.

Figure 9 Changes (L7-M8) in soil moisture capacities (cm) imposed to represent deforestation of tropics.

Figure 10 Differences in precipitation (mm/d) for December to February between experiments (M9-L6) with changes in albedo over tropical forest regions. Areas with increases (decreases) of over 1 mm/d lightly (heavily) shaded.

Figure 11 Time-latitude diagram of differences in convective precipitation (mm/d) for (a) Amazon (b) Congo regions. Areas with decreases shaded. Note that most of the model rainfall in these regions is convective.

Figure 12 Differences in precipitation (mm/d) for June to August between experiments (a) (M9-L6), (b) (M8-L7) simulating tropical deforestation. Areas with decreases shaded.

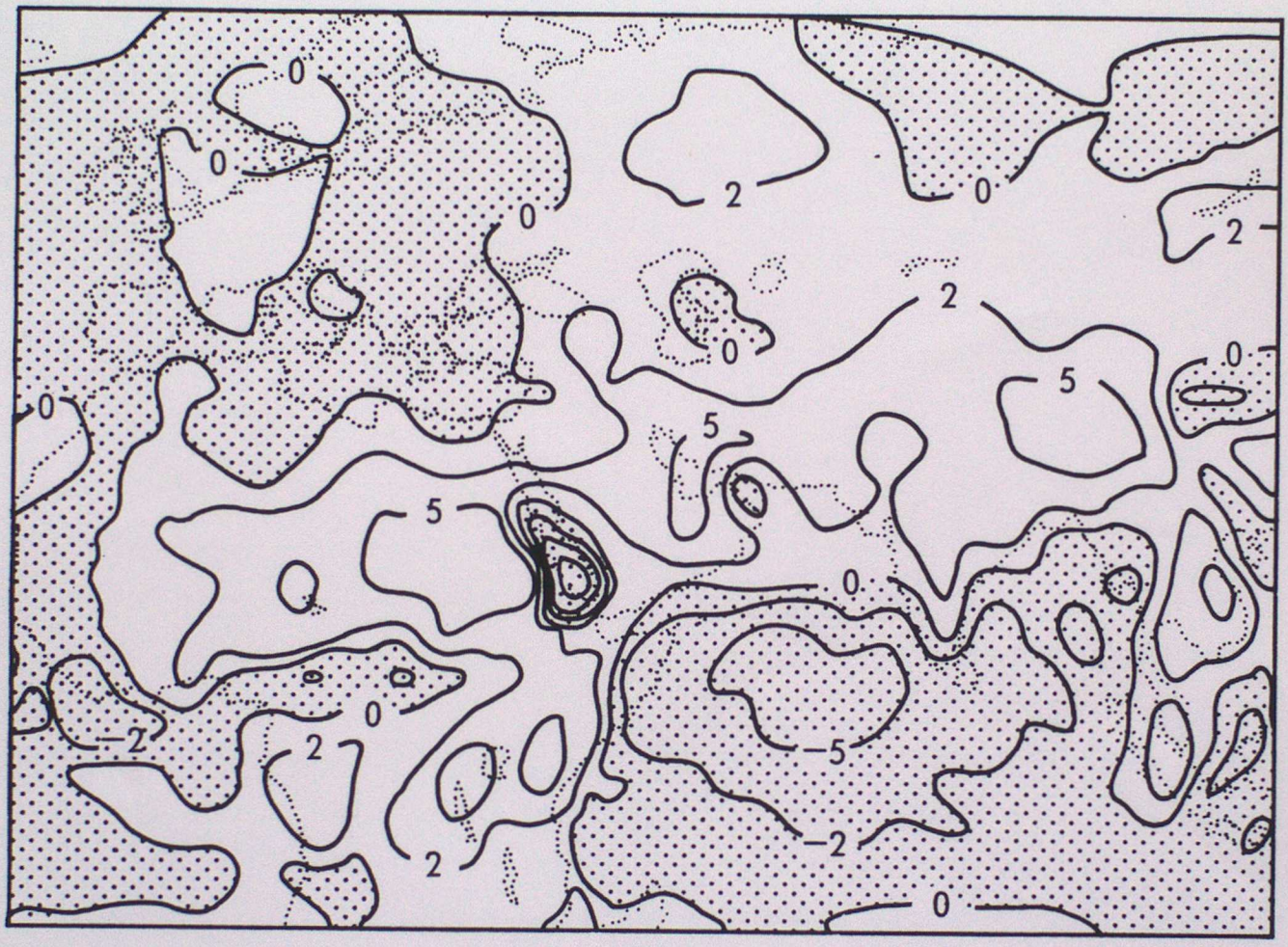


Figure 1 (a) Excess of precipitation with albedo of 0.1 relative to that with albedo of 0.3 at Days 21-110 (soil moisture fixed at 15 cm) in experiments with the Meteorological Office low resolution 5-layer model. Areas with decreases shaded.

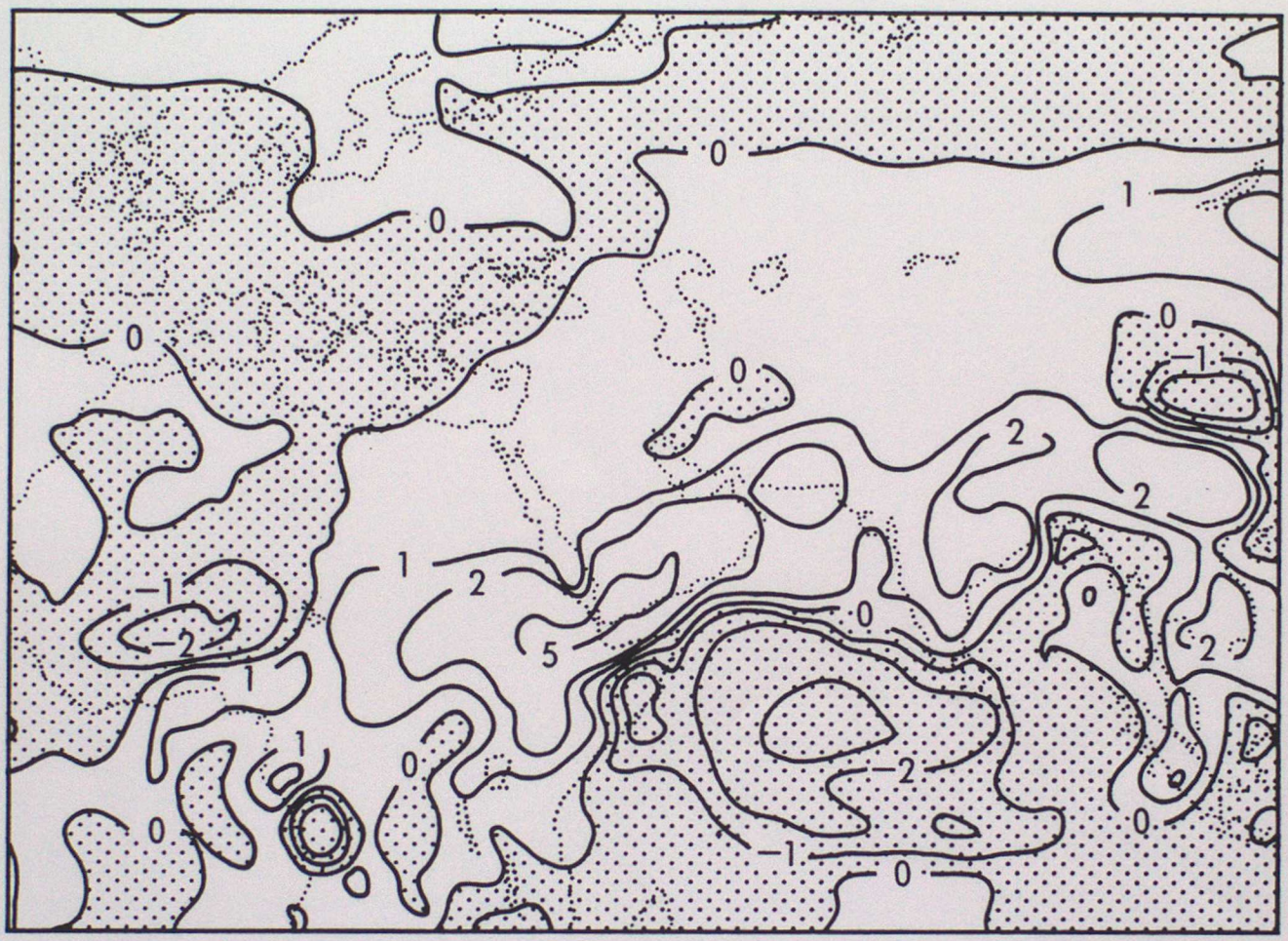


Figure 1 (b) As (a) but for precipitation with albedo of 0.2 relative to that with albedo of 0.3 with interactive initially zero soil moisture.

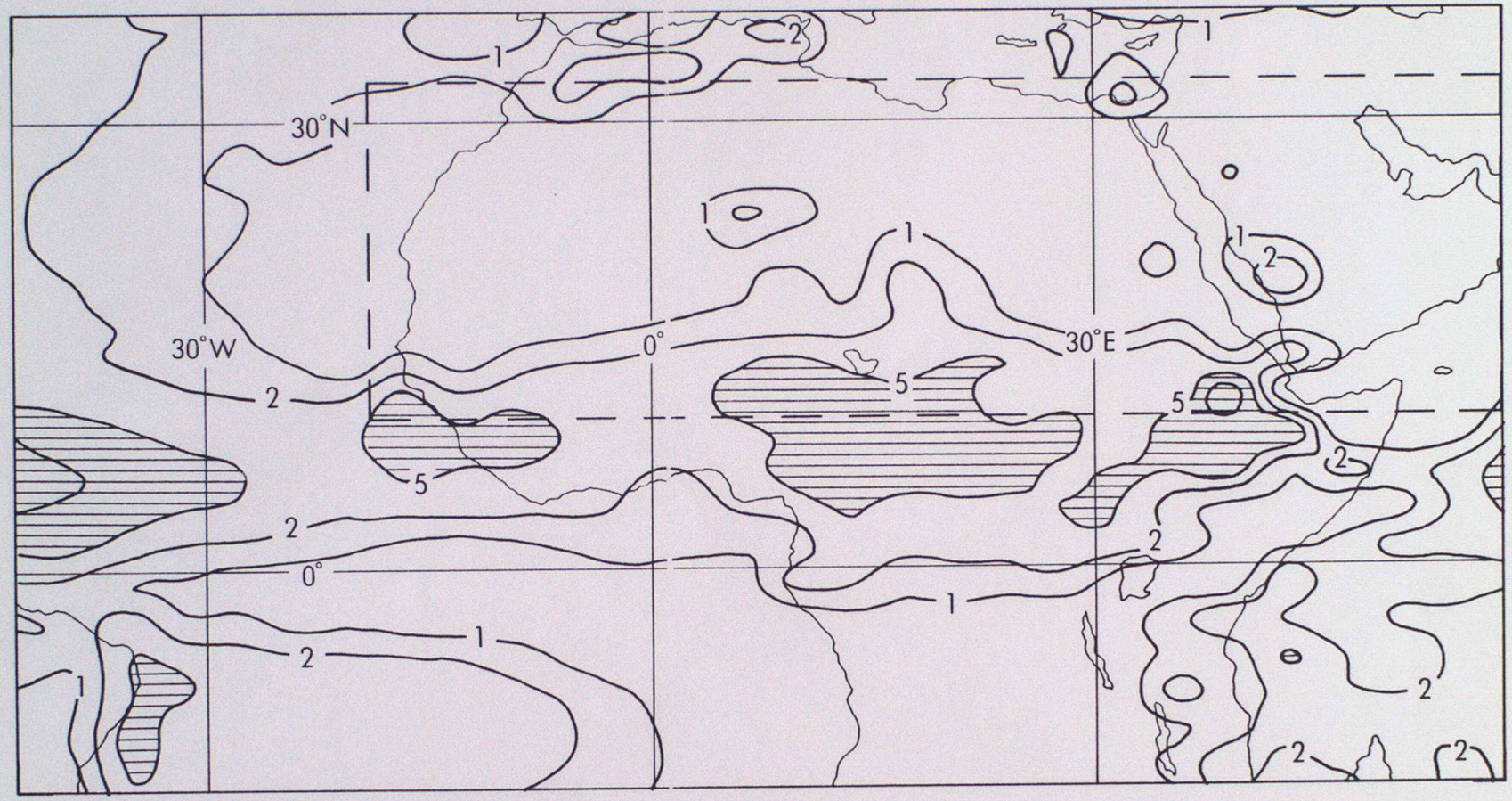


Figure 2 (a)

Figure 2 Precipitation (mm/d) averaged for (a) June to August, (b) September to November of Years 1 to 4 of the control experiment which started on July 24 of Year 0. Contours at 1,2,5, 10 mm/d. Areas with over 5 mm/d shaded. The area with reduced moisture capacity is indicated in Fig 2(a) (dashed).

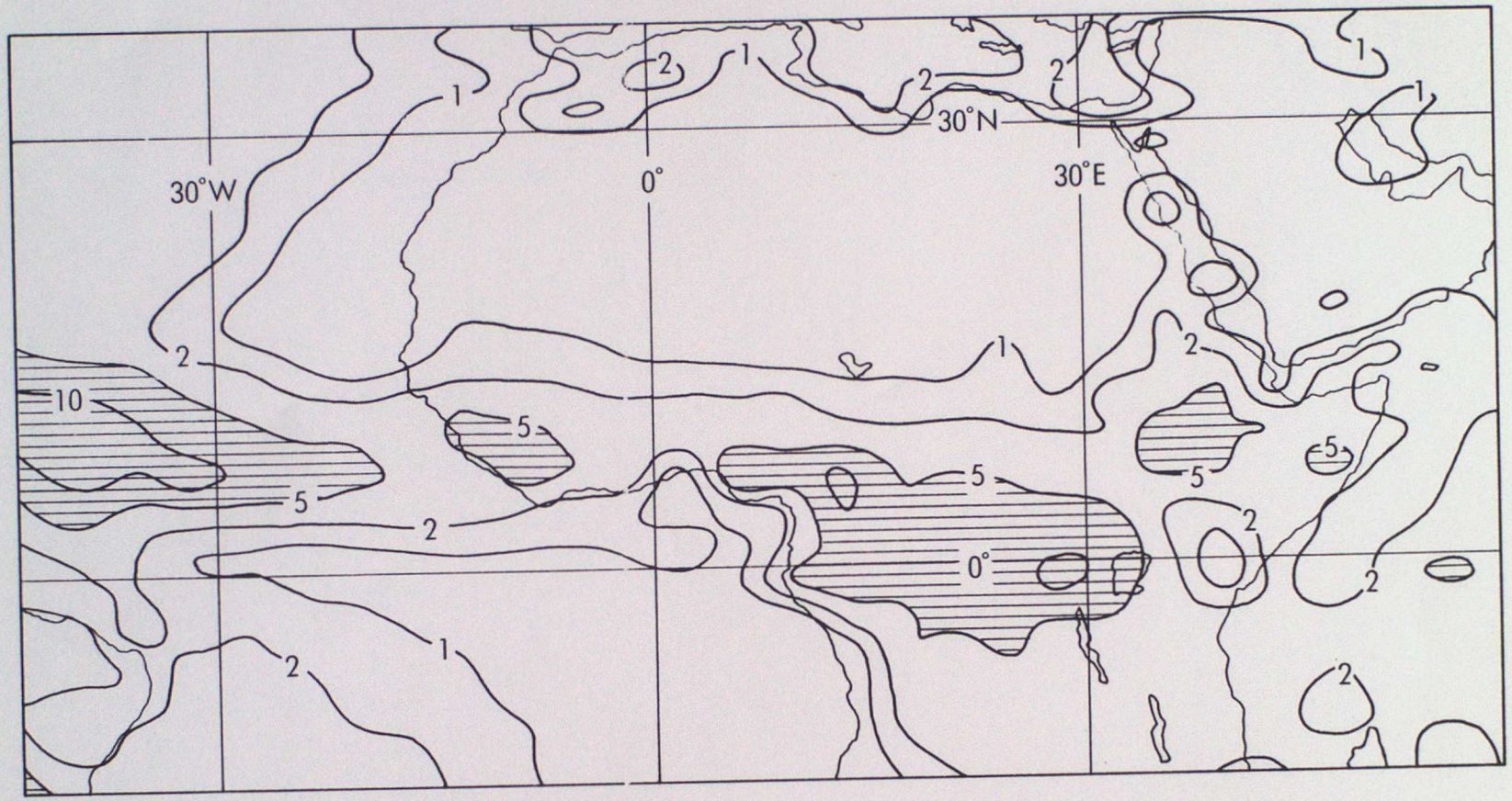


Figure 2 (b)

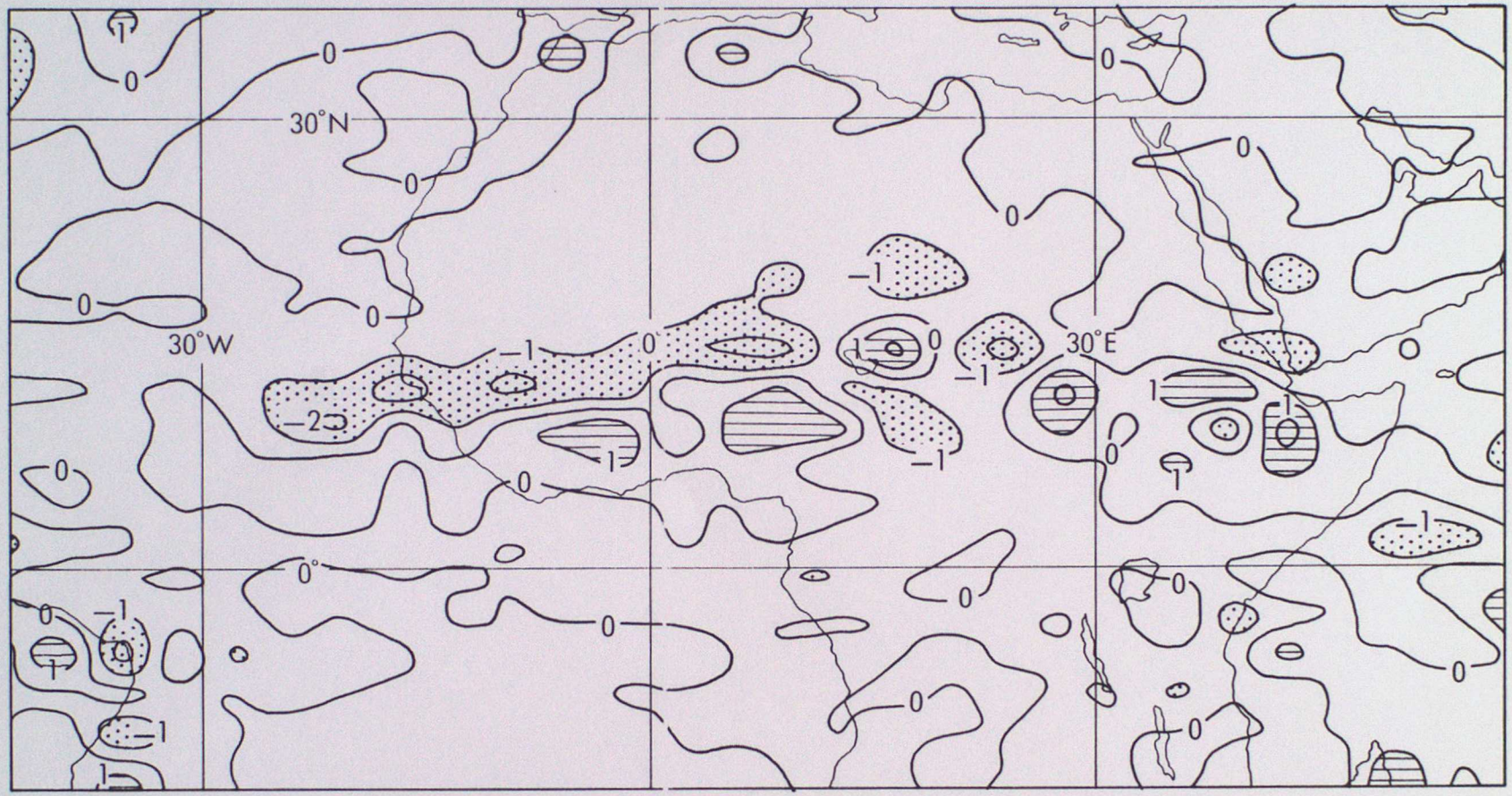


Figure 3 (a)

Figure 3 Differences in rainfall (mm/d), relative to the control average (Fig 2) for June to August of experiments with reduced soil water capacity starting from March 1 of (a) Year 2 and (b) Year 3 of the control experiment. Contours at 0, ±1,2,5 mm/d with increases (decreases) of over 1 mm/d hatched (stippled). Dashed lines in (b) outline the area significant at the 10% level with a 2-tailed Student's t-test.

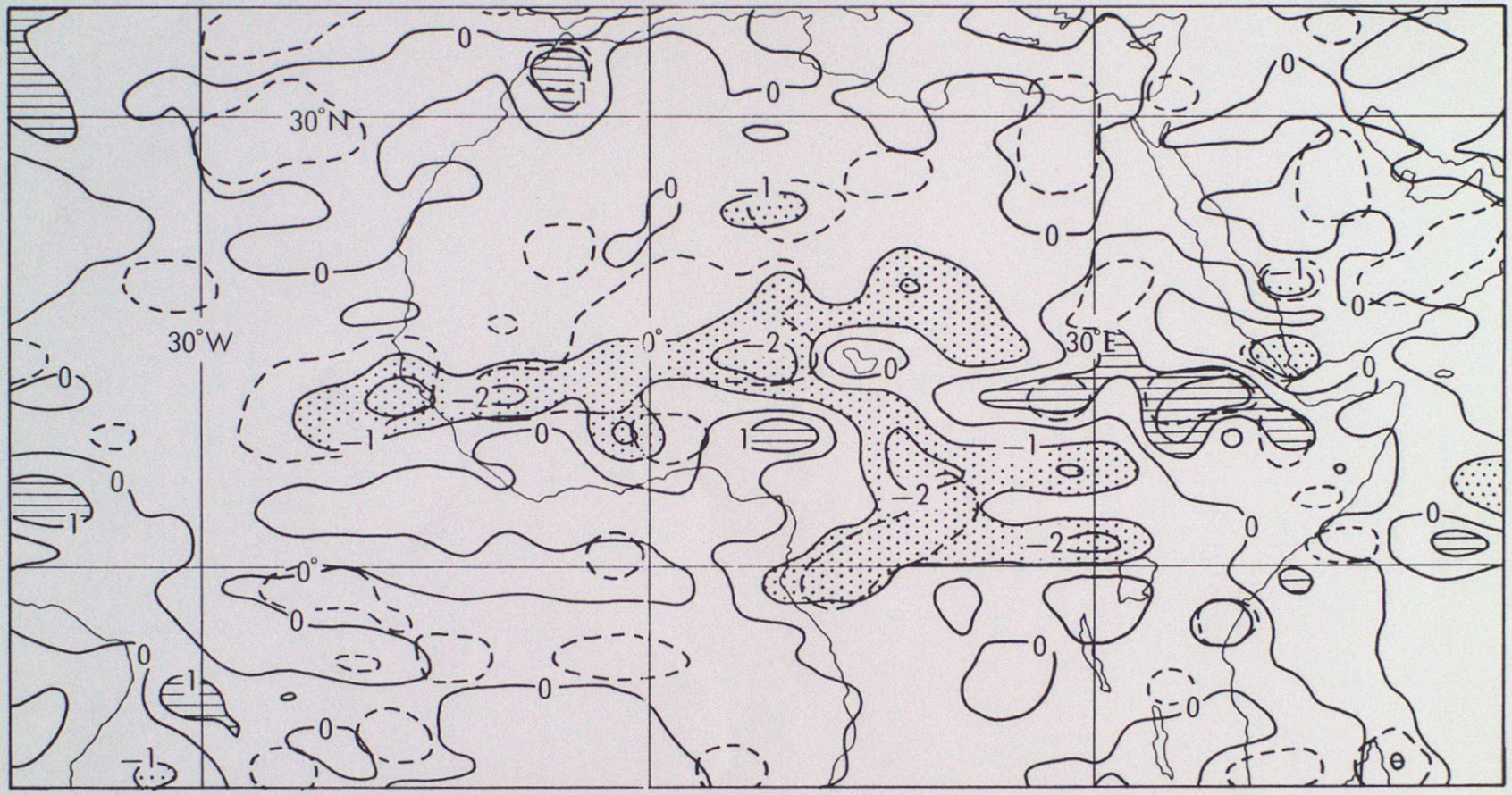
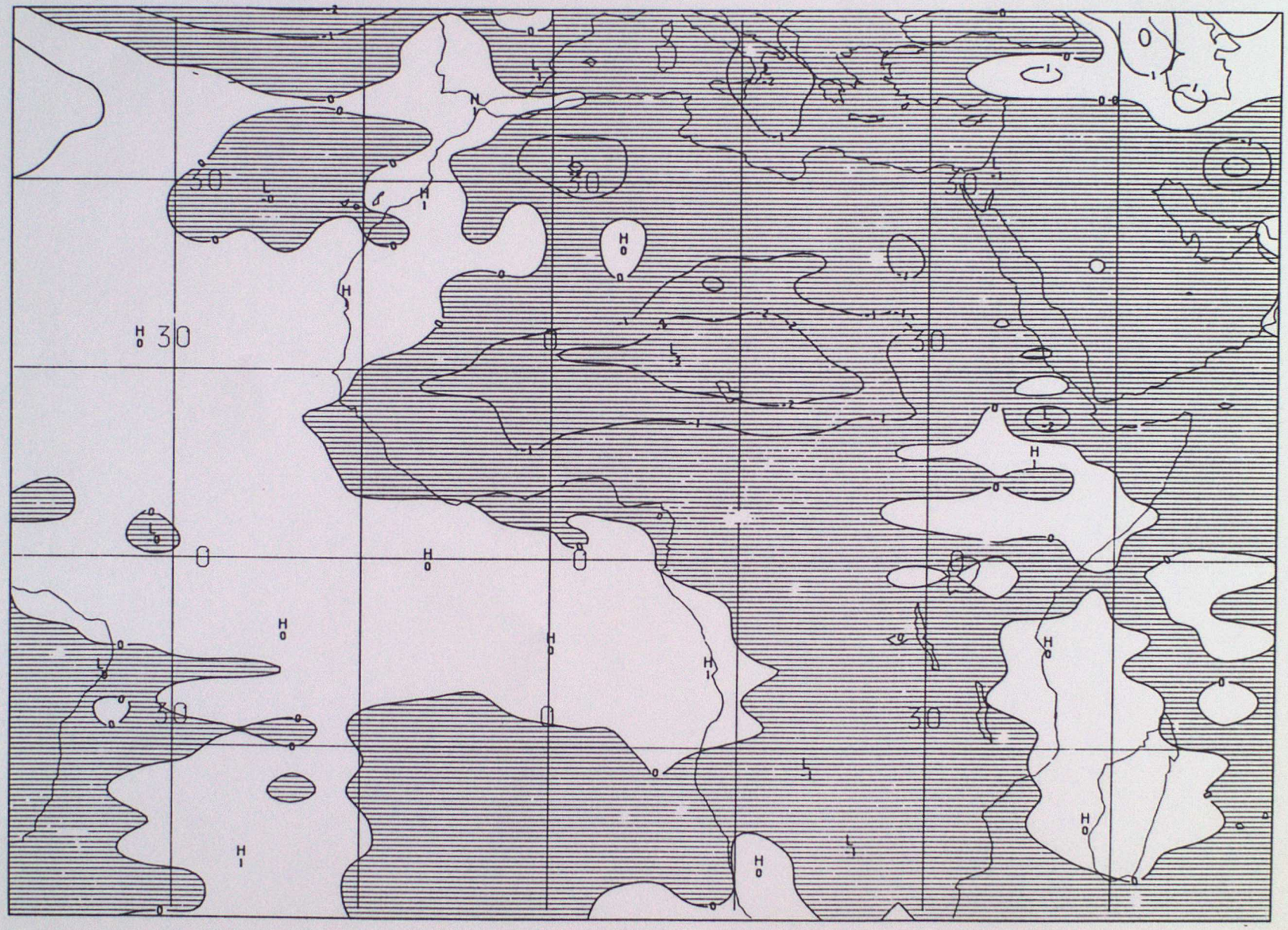


Figure 3 (b)

Figure 4 Differences in sea level pressure (mbar) relative to the control average for June to August of experiment with reduced soil water capacity starting from March 1 of Year 3 of the control experiment.



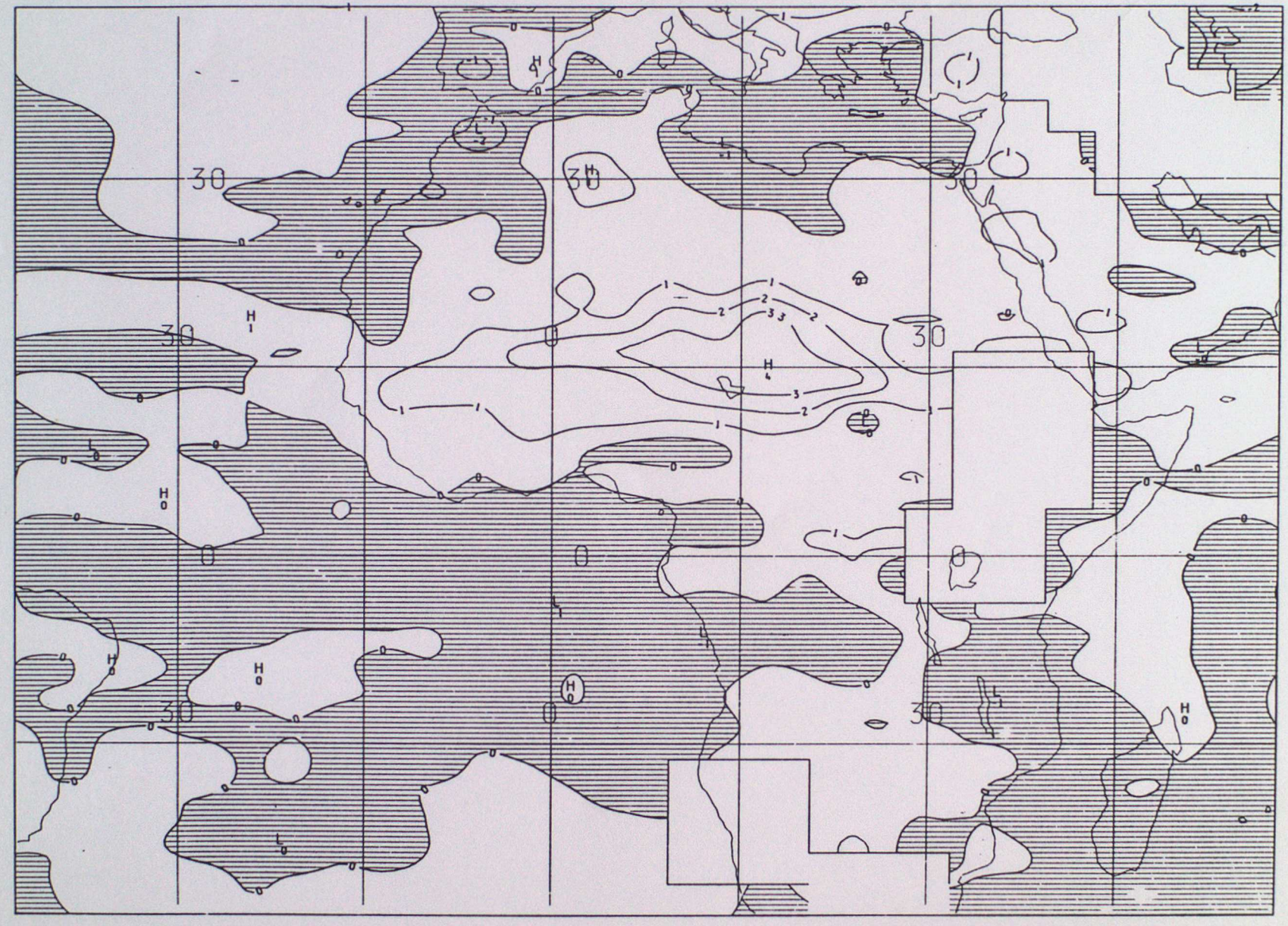


Figure 5 As Figure 4 but for 850 mbar temperature (K).

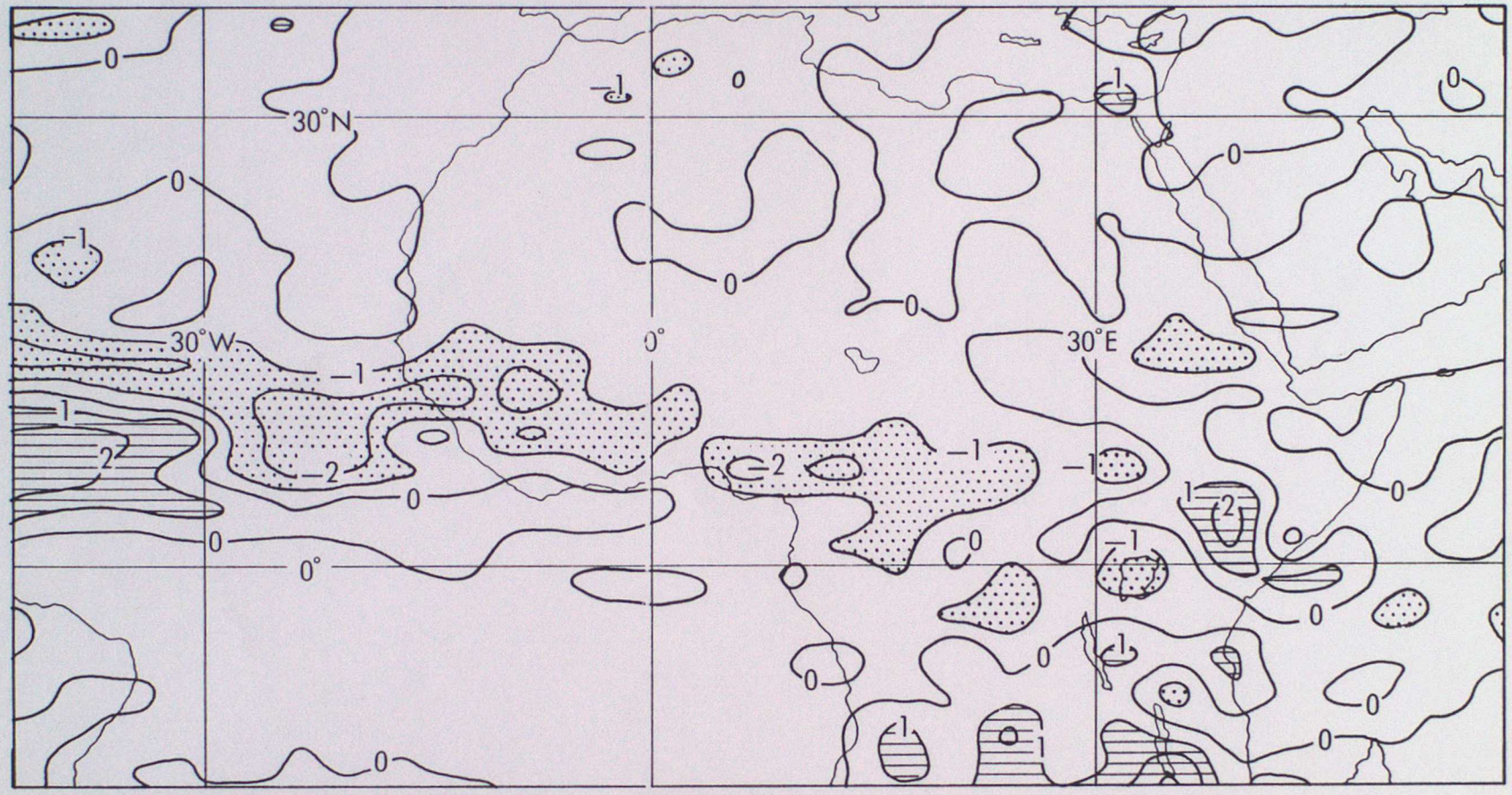


Figure 6 (a)

As Figure 3 (a) but for September to November

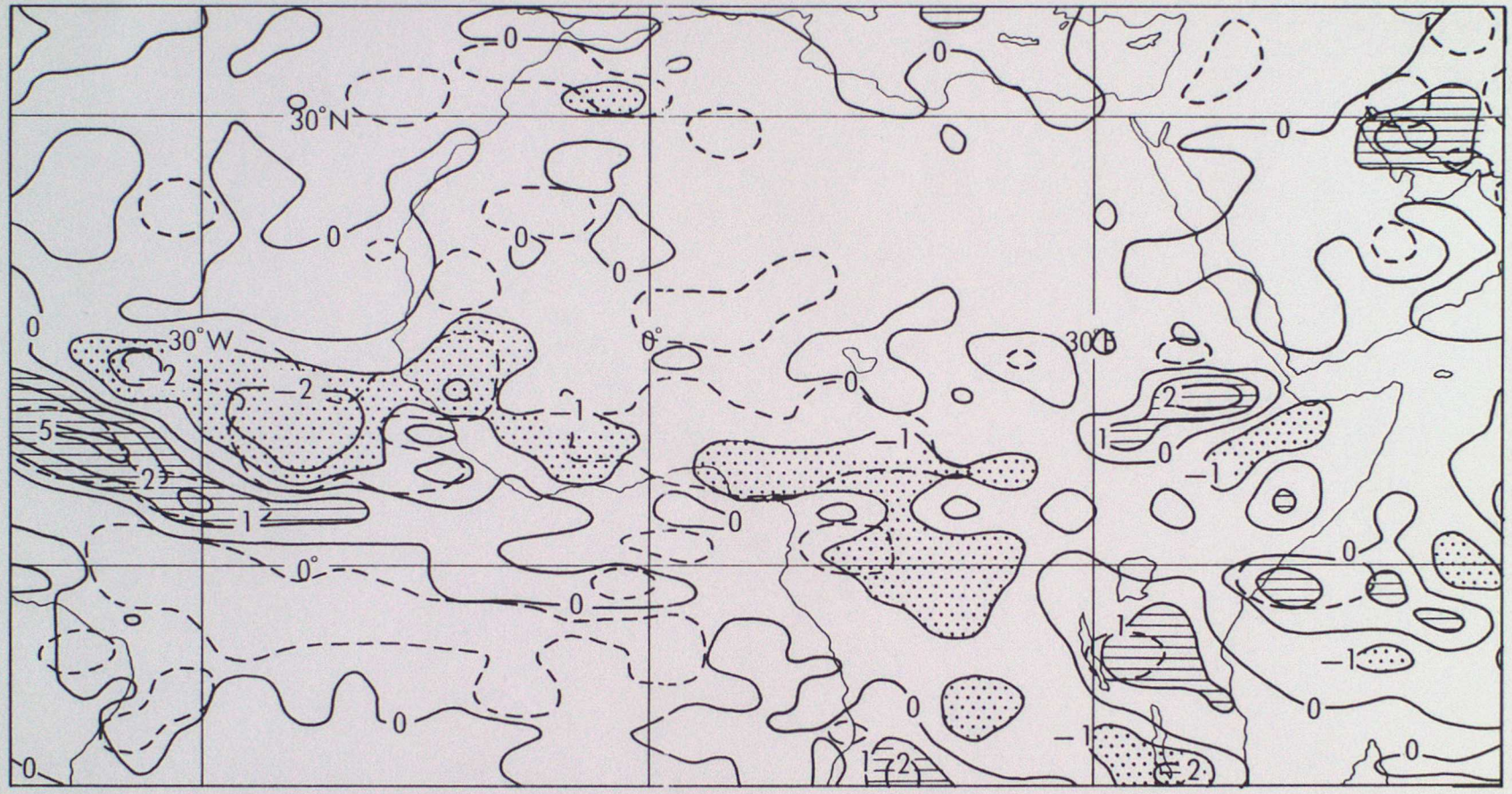


Figure 6 (b)

As Figure 3 (b) but for September to November

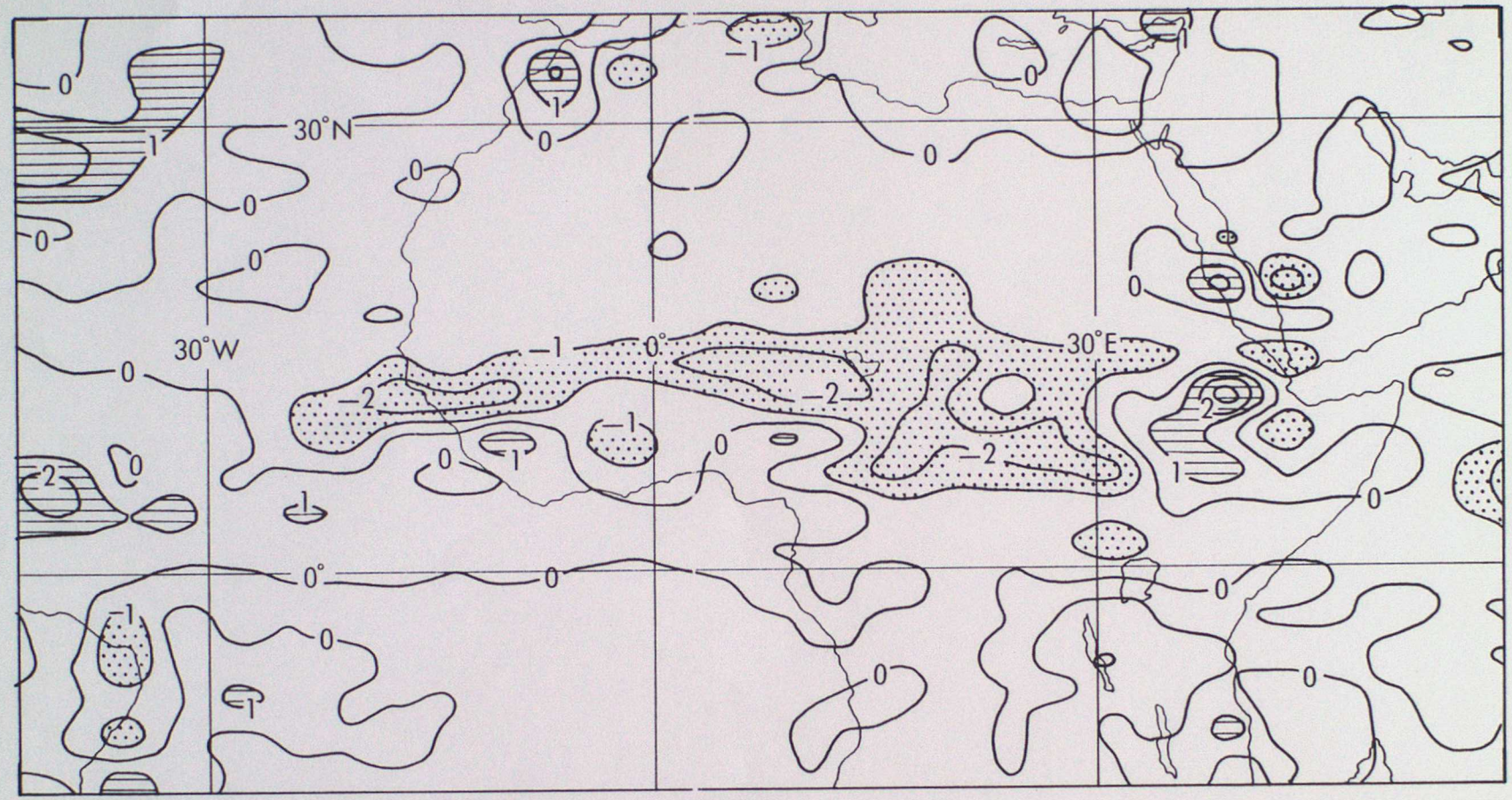


Figure 7 As Figure 3(b) for experiment with reduced soil moisture capacity and increased albedos. (No t-test results)

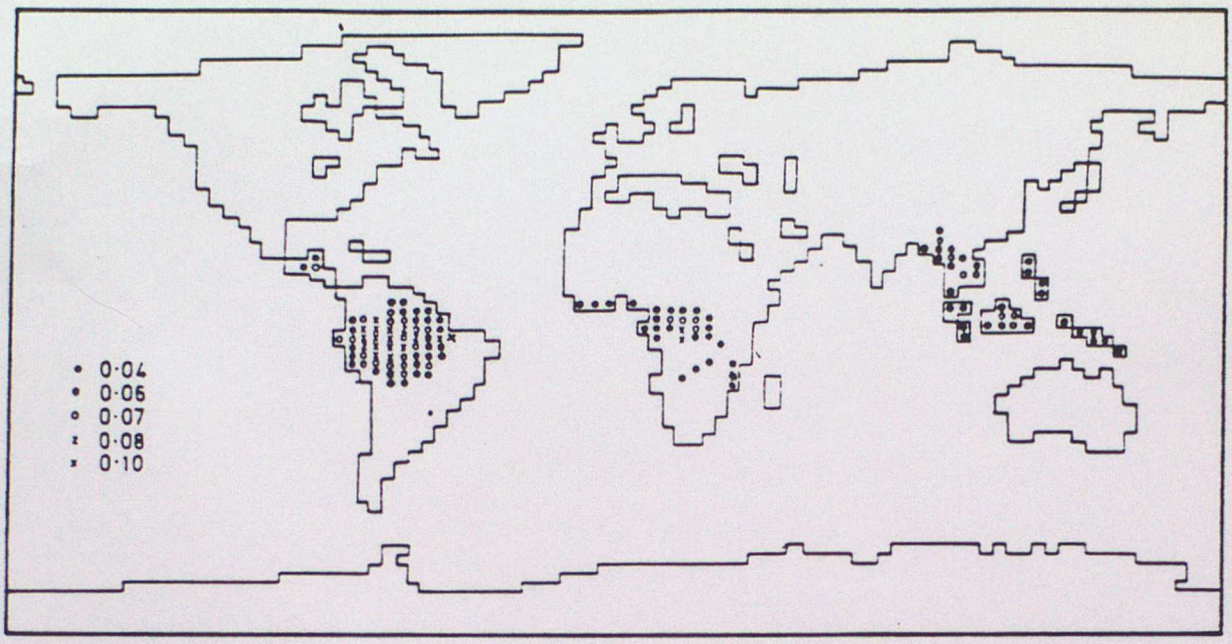


Figure 8 Changes (M8-L7, M9-L6) in albedos imposed to represent deforestation of tropics.

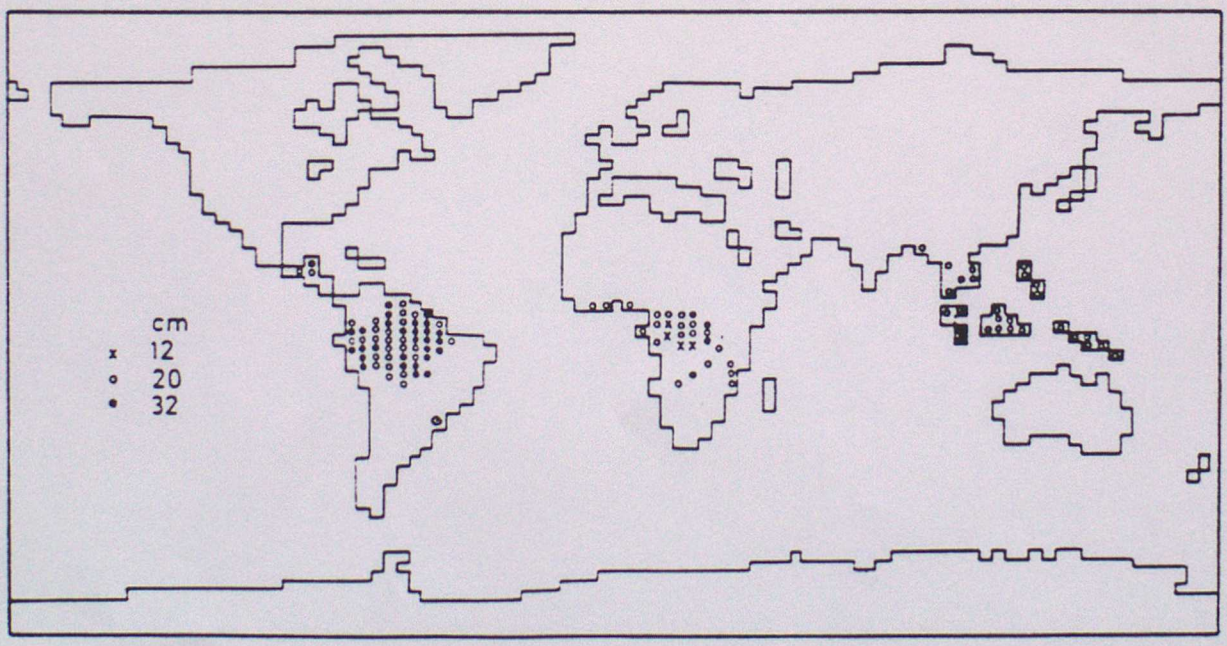


Figure 9 Changes (L7-M8) in soil moisture capacities (cm) imposed to represent deforestation of tropics.

Figure 10 Differences in precipitation (mm/d) for December to February between experiments (M9-L6) with changes in albedo over tropical forest regions. Areas with increases (decreases) of over 1 mm/d lightly (heavily) shaded.

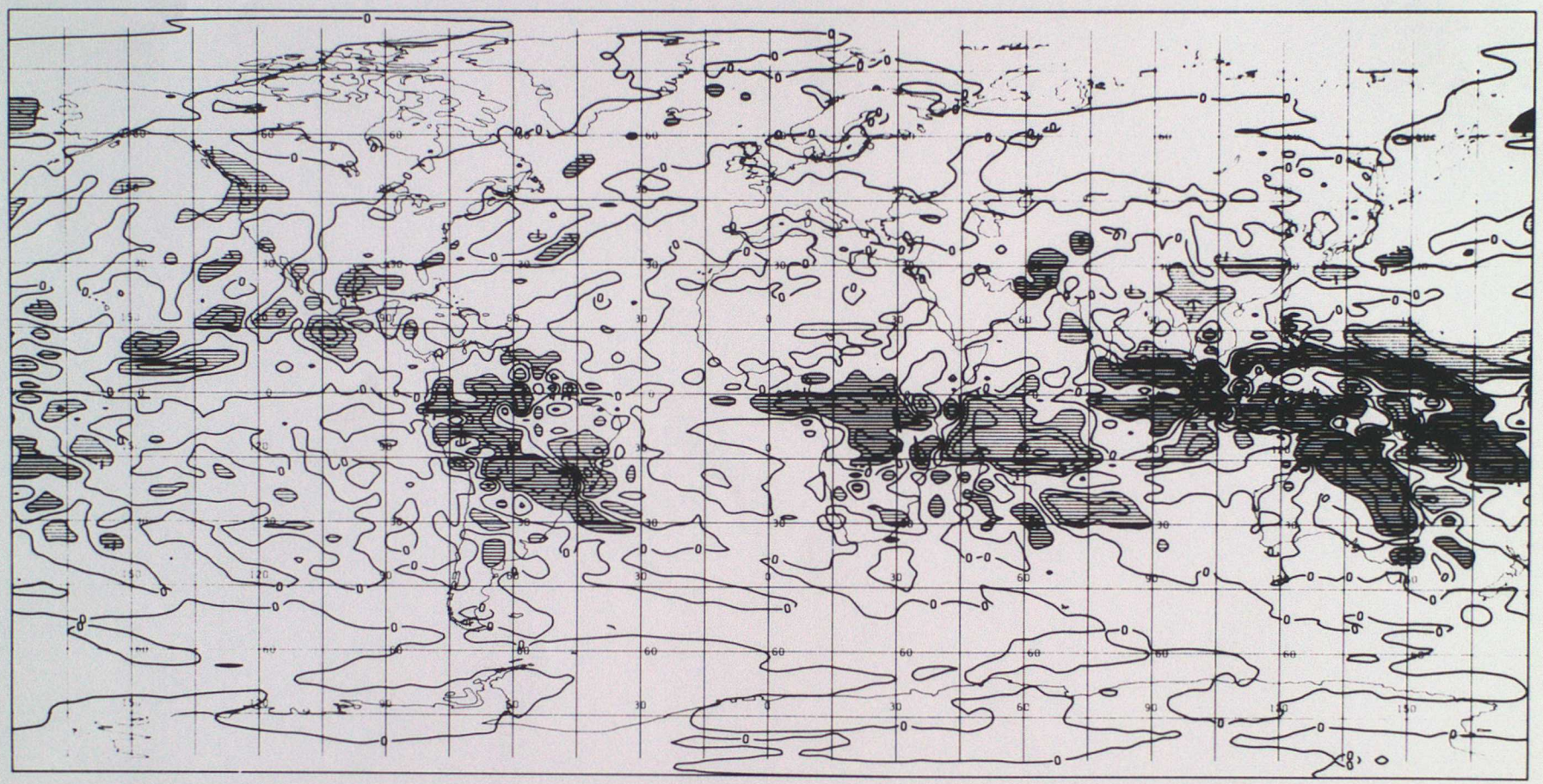


Figure 11 (a)

Figure 11 Time-latitude diagram of differences in convective precipitation (mm/d) for (a) Amazon (b) Congo regions. Areas with decreases shaded. Note that most of the model rainfall in these regions is convective.

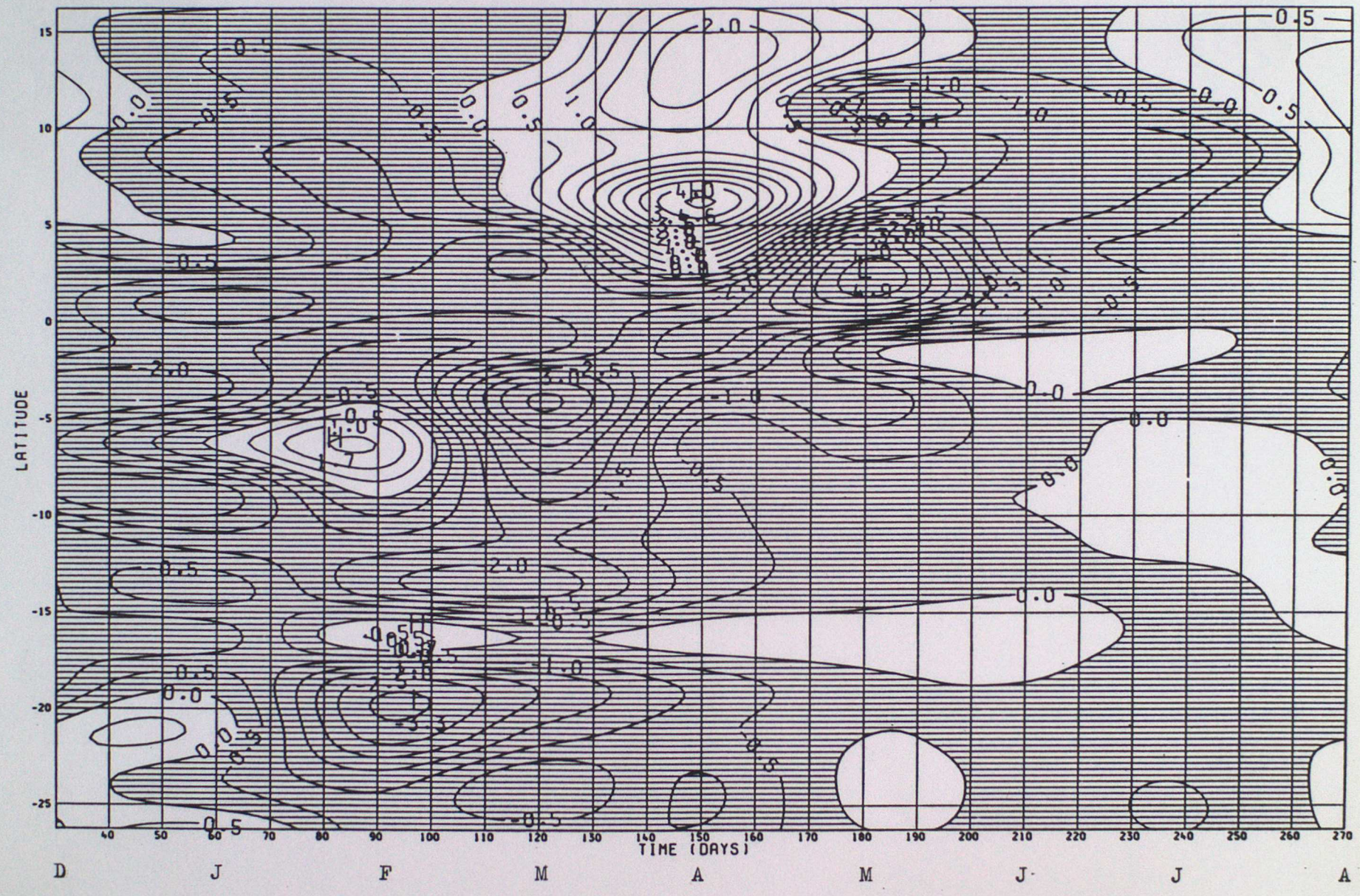


Figure 11 (b)

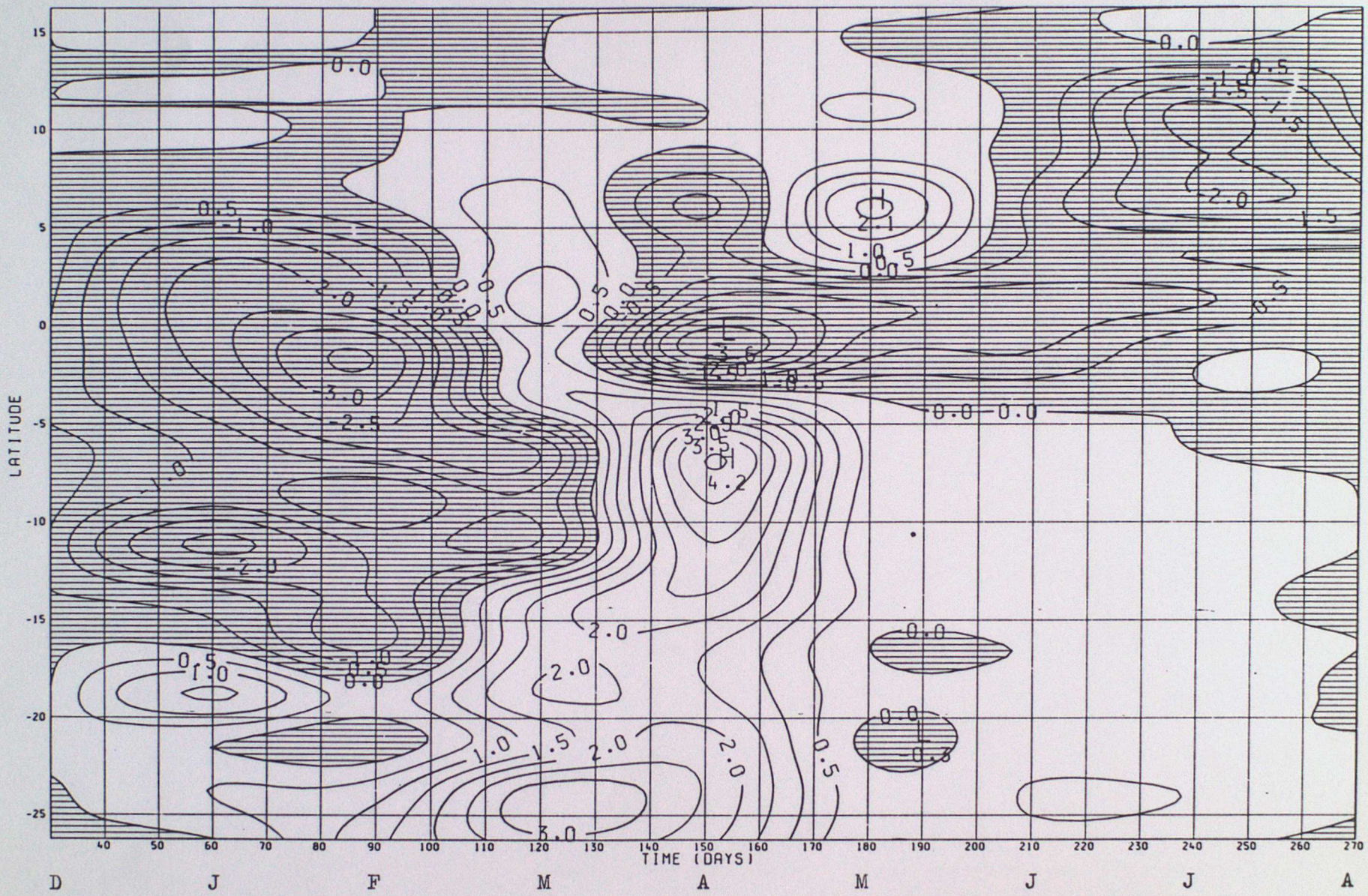

LATITUDE
15
10
5
0
-5
-10
-15
-20
-25
40 50 60 70 80 90 100 110 120 130 140 150 160 170 180 190 200 210 220 230 240 250 260 270
TIME (DAYS)
D J F M A M J J A

Figure 12 (a)

Figure 12 Differences in precipitation (mm/d) for June to August between experiments (a) (M9-L6), (b) (M8-L7) simulating tropical deforestation. Areas with decreases shaded.

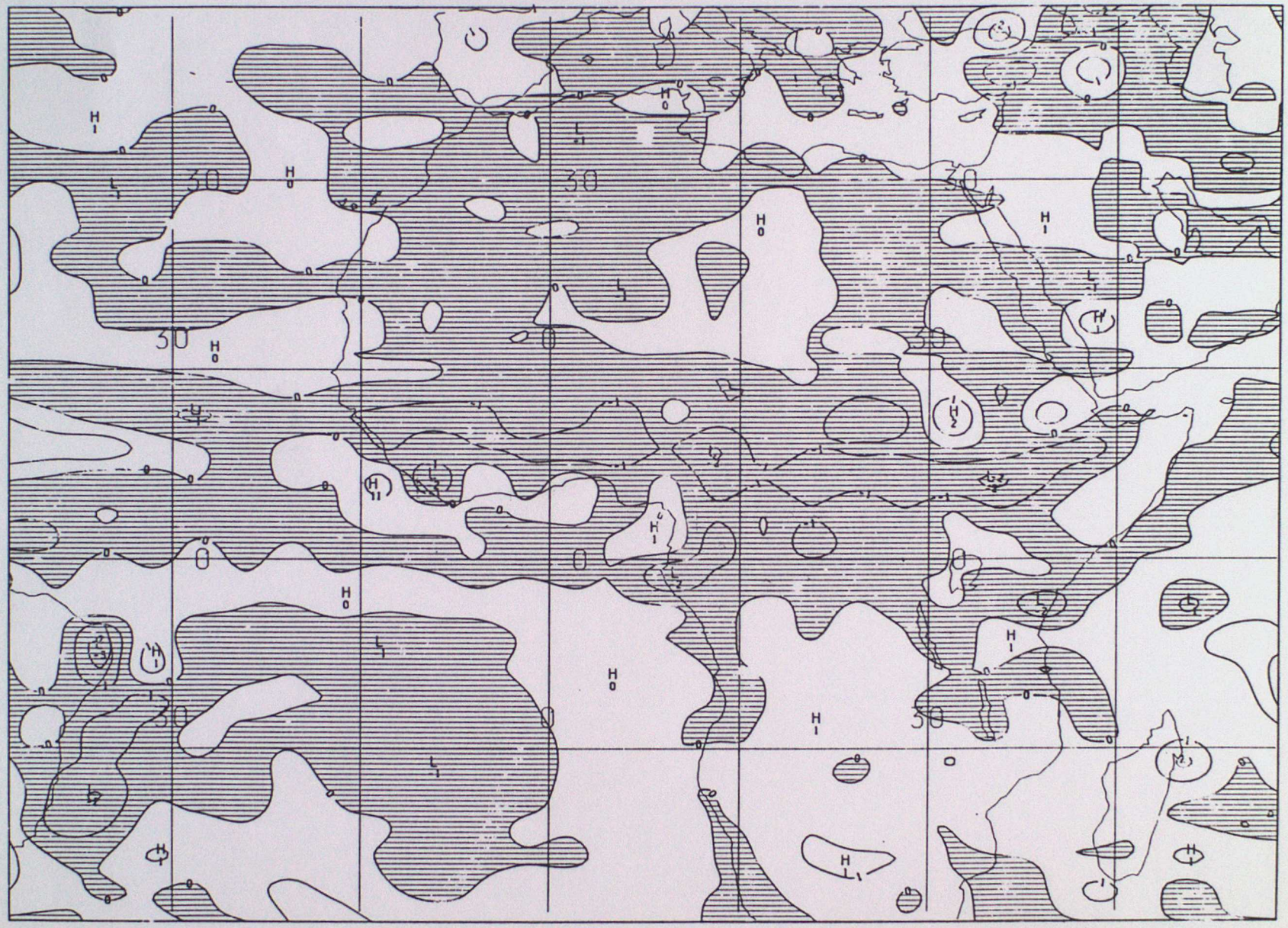
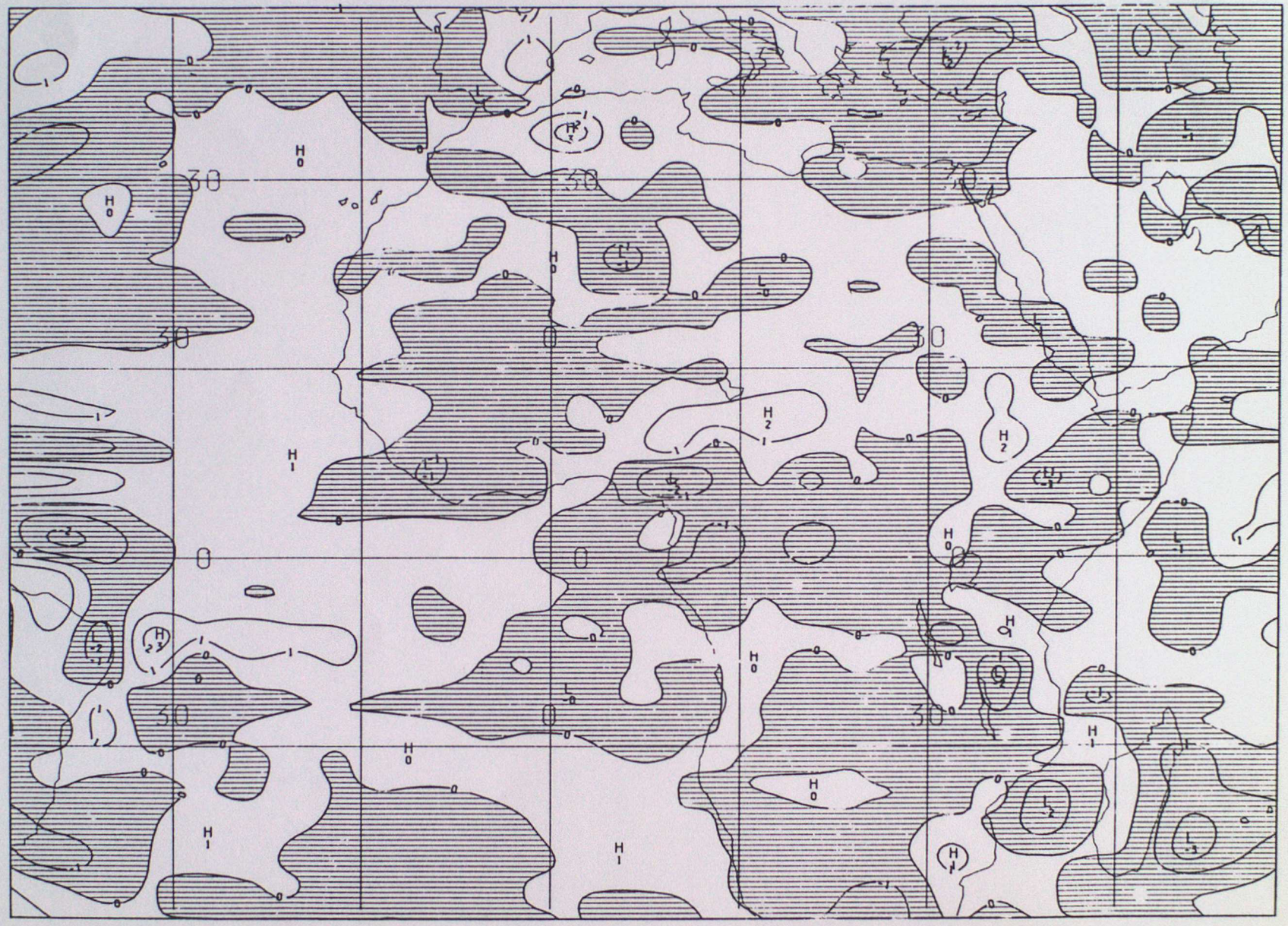


Figure 12 (b)



DYNAMICAL CLIMATOLOGY TECHNICAL NOTES

DCTN 1	June 1984	A scheme for incorporating the thermodynamic sea-ice model into a coupled ocean-atmosphere model. C Gordon and M Bottomley
DCTN 2	Sept 1984	The parametrization of the upper ocean mixed layer in coupled ocean/atmosphere models. C Gordon and M Bottomley
DCTN 3	Sept 1984	The response of a simple thermodynamic sea-ice model to GCM forcing appropriate for perennial sea-ice. M Bottomley
DCTN 4	Sept 1984	Potential vorticity in the stratosphere derived using data from satellites. S A Clough, N S Grahame and A O'Neill
DCTN 5	Sept 1984	Simulation of the seasonal cycle of SST using globally specified climatological data to force simple ocean models. N S Grahame
DCTN 6	Sept 1984	Development of an improved GCM radiation scheme. A Slingo and R C Wilderspin
DCTN 7	Sept 1984	On the specification of surface fluxes in coupled atmosphere/ocean general circulation models. J F B Mitchell, C A Wilson and C M Price
DCTN 8	Sept 1984	A survey of possible repercussions of the Soviet river reversal programme on Arctic sea-ice. H Cattle
DCTN 9	Oct 1984	A study of a strong stratospheric warming during January 1982 T D A Fairlie
DCTN 10	Nov 1984	An intercomparison of the Met O 11 N45 model and the Met O 20 11-level model on the Cyber 205 J F Dyson
DCTN 11	Nov 1984	Simulation of observed time series of temperature and precipitation over eastern England. D N Reed
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DCTN 13	Nov 1984	On modelling the effects of CO_2 on climate. J F B Mitchell
DCTN 14	Dec 1984	Development of a revised longwave radiation scheme for an atmospheric general circulation model. A Slingo and R C Wilderspin
DCTN 15	Jan 1985	Rossby wavetrains in the stratosphere forced by localised disturbances in the troposphere. C J Marks (Oxford), A O'Neill and V D Pope

DYNAMICAL CLIMATOLOGY TECHNICAL NOTES

DCTN 16	Jan 1985	Observing system experiments using the Meteorological Office's 15-level model. R A Bromley
DCTN 17	Feb 1985	Atmospheric general circulation simulation of the climate of the Mediterranean region. H Cattle
DCTN 18	Feb 1985	Diurnal variation and cloud in a general circulation model. C A Wilson and J F B Mitchell
DCTN 19	Feb 1985	The sensitivity of the Saharan region in a general circulation model. W M Cunnington and P R Rowntree
DCTN 20	Mar 1985	The impact of aircraft observations on analyses and forecasts from the operational 15-level model. B R Barwell
DCTN 21	Mar 1985	Non-linear effects on heat-induced circulation of the tropical atmosphere. A E Gill and P J Phlips
DCTN 22	Apr 1985	Simulation of cloudiness with the U.K. Meteorological Office 11-layer AGCM. A Slingo and R C Wilderspin
DCTN 23	Apr 1985	An integrated approach to the representation of cloud processes. R N B Smith
DCTN 24	May 1985	Nuclear Winter: Calculations of the shortwave radiative effects of soot aerosols. A Slingo and P Goldsmith
DCTN 25	Jul 1985	Sea ice and the Antarctic winter circulation: a numerical experiment. J F B Mitchell and T S Hills
DCTN 26	Jul 1985	A Budyko-Sellers model for study of ocean-atmosphere interaction. W J Ingram
DCTN 27	Aug 1985	The effect of a shoaling thermocline on equatorially-trapped Kelvin waves. A E Gill and B A Ming